VOL. 257 JUNE 1959 No. 6

# American Journal of Science

**Established in 1818 by Benjamin Silliman**

EDITORS
JOHN RODGERS—JOSEPH T. GREGORY

ASSOCIATE EDITORS

GEORGE GAYLORD SIMPSON, NEW YORK, N. Y.
A. F. BUDDINGTON, PRINCETON, N. J.
CARL O. DUNBAR
HENRY MARGENAU
HORACE WINCHELL
EDWARD S. DEEVEY, JR.
MATT WALTON
RICHARD F. FLINT
HAROLD G. CASSIDY
G. EVELYN HUTCHINSON
NEW HAVEN, CONN.
WILMOT H. BRADLEY, WASHINGTON, D. C.
LEASON H. ADAMS, LOS ANGELES, CALIF.
HOWEL WILLIAMS, BERKELEY, CALIF.
C. R. LONGWELL, PALO ALTO, CALIF.
ADOLPH KNOPF, STANFORD, CALIF.
FREDERICK J. ALCOCK, OTTAWA, CANADA
CECIL E. TILLEY, CAMBRIDGE, ENGLAND

STERLING TOWER
YALE UNIVERSITY
NEW HAVEN, CONNECTICUT

Published 10 times a year. Ten dollars a year. One dollar twenty-five cents a copy.
Foreign, ten dollars sixty cents Canada, ten dollars thirty-five cents.

# AMERICAN JOURNAL OF SCIENCE

*Established in 1818 by Benjamin Silliman*

## THE FIRST SCIENTIFIC JOURNAL IN THE UNITED STATES

## DEVOTED TO THE GEOLOGICAL SCIENCES AND TO RELATED FIELDS

*Editors:* JOHN RODGERS—JOSEPH T. GREGORY
*Managing Editor:* BETH A. BROMBERT

Entered as second-class matter at the Post Office at New Haven, Conn., under the Act of March 3, 1879. Published monthly, except July and September, at Sterling Tower, Yale University, New Haven, Conn.

*Subscription rate* $10 per year ($1.25 a number). Student rate $5 per year (application blanks available on request). Postage prepaid to the United States; 35 cents per year to Canada; 60 cents per year elsewhere.

The JOURNAL completed its *first* series of 50 volumes in 1845, its *second* series of 50 volumes in 1870, its *third* series of 50 volumes in 1895, its *fourth* series of 50 volumes in 1920, its *fifth* series of 36 volumes in 1938. Since 1938 the numbering by series has been discontinued, and volumes are annual and bear the whole numbers as of 1818, that for 1939 being volume 237.

Back numbers and volumes, also cumulative indices of all series, may with few exceptions be obtained from the office of the AMERICAN JOURNAL OF SCIENCE, Sterling Tower, New Haven, Conn. Positive microfilm copies of each complete volume may be purchased by subscribers to the paper edition from *University Microfilms*, 313 N. First St., Ann Arbor, Michigan. Regular subscribers may purchase each past year in microcard form, from *The Microcard Foundation*, Wesleyan University, Middletown, Connecticut.


### INSTRUCTIONS TO CONTRIBUTORS

*Typescripts* of articles submitted should be double-spaced, should follow the recommendations in the pamphlet *Suggestions to Authors*, 5th ed. (see below), and should be preceded by brief abstracts. They should be checked with special care before being submitted, to avoid changes in proof other than printer's errors.

*References to literature* should be listed alphabetically by authors at the end of the article; all references should be complete. For articles in journals or other serial publications, the reference should give in order: Author's name, year of publication, full title: name of periodical or series, volume number, inclusive pages. For books or other unnumbered publications, it should give: Author's name, years of publication, full title: place of publication, publisher. This form follows the recommendations in the pamphlet "Abbreviations used in publications of the U. S. Geological Survey", which is also the standard used for abbreviating names of periodicals, etc. Citations within an article should be to author and year, with specific pages wherever appropriate.

*Plates* (photographs), *figures* (line drawings), and *footnotes* should each be numbered consecutively through each article, using arabic numerals. If two photographs form one plate, they are figures 1 and 2 of that plate.

*Illustrations* should be neat and legible and should include explanations of symbols used. Copy that cannot be reproduced cannot be accepted. Plates and figures should be capable of reduction to not more than 4 by 7 inches, all lettering being at least 1/16th inch high after reduction. When necessary, one large map or table can be accepted, if it will not exceed 7 inches in width after reduction. Line drawings should be in black India ink on white drawing board, tracing cloth, or coordinate paper printed in blue. Photographs should be positive glossy prints. Photostatic and typewritten material cannot be accepted as copy for illustrations.

*Reprints.* Thirty separate copies of each article will be furnished to the author free of cost and without previous notice from him; these will be without a cover. Additional copies will be furnished at cost, which will of course be greater if the article is accompanied by plates involving unusual expense. Copies will be furnished with a printed cover giving the title, author, volume, page, and year, when specially ordered.

---

Suggestions to Authors of the Reports of the United States Geological Survey, 5th ed. Washington, D. C., 1958 (Government Printing Office, $1.75).

## ANNOUNCEMENT

The American Journal of Science announces the publication of Volume 1 of its Radiocarbon Supplement, to be devoted to the publication of date lists from radiocarbon laboratories all over the world. Professors Richard Foster Flint and Edward S. Deevey, Jr. are the Editors of the Supplement.

Volume 1 contains fourteen date lists and a bibliography of radiocarbon literature, and totals 218 pages. In the future, volumes will appear once a year unless a great increase in the amount of material requires more frequent publication.

The price of Volume 1 will be $2.50 until January 1, 1960 after which the price will be $4.50 per volume. The office of the Supplement is the same as that of the American Journal of Science, Box 1905A, Yale Station, New Haven, Connecticut.

John Rodgers
Joseph T. Gregory
Editors, American Journal of Science

[American Journal of Science, Vol. 257, June 1959, P. 385-407]

American Journal of Science

*JUNE 1959*

# THE ORIGIN OF DEFORMATION LAMELLAE IN QUARTZ

JOHN M. CHRISTIE and C. B. RALEIGH

Department of Geology, University of California, Los Angeles, California

ABSTRACT. Fabrics of four quartzite specimens containing numerous quartz grains with deformation lamellae are described in detail. Patterns of preferred orientation of deformation lamellae in all four specimens are similar in that the poles of the lamellae define a small-circle girdle (about an axis designated A). The orientations and strengths of maxima within the girdle, however, are not consistent in different specimens. [0001]-axes of grains containing deformation lamellae also define a small-circle girdle about the same axis (A). In each specimen the great-circles containing [0001] and the pole of the deformation lamellae for individual grains pass through, or close to, the axis A of the small-circle girdles.

The deformation lamellae are shown to be late structures unrelated to the deformations which induced the preferred orientation of the quartz grains in the rocks. The lamellae are not parallel to rational crystallographic planes and they are considered to represent kink-bands resulting from shearing parallel to [0001] on irregular planes in the zone of [0001]. The shearing is probably controlled by imperfections in the crystal structure, which commonly exist parallel to [0001] in quartz.

According to this hypothesis the axis (A) of the small-circle girdle defined by poles of lamellae is the axis of maximum compressive stress during the deformation which produced the lamellae. This relationship may be used to obtain a dynamic interpretation of deformation lamellae in quartzose sedimentary and metamorphic rocks. The hypothesis is tested using data from the Baraboo Quartzite and it is demonstrated that the deformation lamellae in the rocks of this formation may be related to the folding of the Baraboo syncline.

## HISTORICAL REVIEW

An extensive literature now exists on the nature and occurrence of lamellar structures in quartz grains in rocks of various types. There appear to be at least two or three different types of lamellae and there is considerable disagreement as to the character and the mode of origin of the lamellae.

Planar structures consisting of minute closely-spaced inclusions were first reported by Böhm (1883). Many writers have since described lamellar structures in the quartz grains of tectonites and they have frequently been designated "Böhm lamellae" or "Böhm striae" (Becke, 1892; Mügge, 1896; Sander, 1930). Intensive studies of the crystallographic orientation of these lamellae and their orientation in rock fabrics have been made by Sander (1930), Hietanen (1938), Fairbairn (1941), Ingerson and Tuttle (1945), Riley (1947), and Savul (1948). Becke (1892) described lamellae in quartz grains in a gneiss from the Central Alps; they were composed partly, but not entirely, of inclusions and showed a refractive index lower than that of the remainder of the crystal. Fairbairn (1941) also stated that the refractive index of lamellae in some cases appears to be lower than that of the grains in which they occur. Ingerson and Tuttle (1945), on the other hand, have described lamellae which are not composed of inclusions, but have refractive in-

dices which are higher than those of the host crystals. They also recognize two other types—one consisting of planes of brown, sometimes liquid, inclusions, also showing a difference of refractive index, and the other consisting only of brownish granular material; they consider the latter to be 'relict' lamellae which have evolved from lamellae of the first type. Since many of the lamellae cannot be resolved microscopically into aggregates of discrete inclusions the practice of refering to all closely-spaced lamellae as 'Böhm lamellae' has been discontinued by many writers.

The lamellae are invariably found in quartz grains which show appreciable post-crystalline strain and it is now established that they are produced by deformation (Fischer, 1925; Fairbairn, 1941; Ingerson and Tuttle, 1945). There are several hypotheses as to the genesis of the lamellae: 1) Becke (1892) considered them to represent partially healed fractures in the quartz grains; 2) Judd (1888) and others, have suggested that they represent secondary twin-lamellae; 3) Mügge (1896), Fischer (1925), Sander (1930), Hietanen (1938), Fairbairn (1941), and Savul (1948) have all maintained that the lamellae are produced by translation-gliding; 4) Ingerson and Tuttle (1945) and Turner (1948) consider that they are microscopic shear-surfaces which are not parallel to rational planes in the quartz lattice.

A number of translation-mechanisms have been postulated to account for the deformation lamellae and there is a notable lack of unanimity among proponents of the translation-gliding hypothesis as to the actual glide-systems involved. Translation-gliding was first proposed by Mügge (1896) who showed that in basal sections of euhedral quartz crystals the traces of lamellae were neither parallel nor perpendicular to the sides of the hexagonal section, suggesting that they are not rational crystallographic planes; he, accordingly, considered that the translation took place on planes subparallel to the base. Sander, in his classic work of 1911, also maintained that the lamellae represent planes of translation-gliding. Fischer (1925) favored the view that translation takes place on {0001}, {10$\bar{1}$1}, {01$\bar{1}$1} and perhaps more obtuse rhombohedra. Although Schmidt (1927) also postulated translation on {0001}, {10$\bar{1}$1}, and {01$\bar{1}$1}, he believed that visible lamellae need not develop parallel to the glide-planes. Sander later (1930) showed, with the aid of the universal stage, that the lamellae are inclined to {0001} at angles varying from 6° to 30° and concluded that the planes of translation are {01$\bar{1}$3} and {01$\bar{1}$2}.

Hietanen (1938) made a detailed study of the orientation of lamellae in different portions of individual grains with undulose extinction from the Finnish quartzites. She postulated several stages in the deformation of quartz grains. Initially, the grains undergo a limited amount of translation-gliding on {0001}, combined with bending of the glide-plane, which gives rise to feeble undulose extinction. In the next stage there is weak deformation of the lattice and the deformation lamellae are formed more or less parallel to the basal plane, the cavities in the striations being produced by breaking of the quartz-lattice. In a still later stage actual fractures are produced more or less parallel to the [0001]-axis and there is gliding parallel to the prism planes. Hietanen considered that some lamellae may also be relics of rhombohedral translation planes. But since the lamellae are never parallel to either the basal

or the unit rhombohedral planes she postulated that they retained their initial orientation while the lattice was reoriented by a later deformational process in which one of the directions of closest packing of Si atoms in the grain lattices becomes oriented parallel to the *a* kinematic axis of the deformation. Hietanen also suggested that the early translation might take place on planes which diverge slightly from {0001}, in view of the 'screw-like' lattice structure of quartz.

Fairbairn (1941) and Ingerson and Tuttle (1945) have made extensive studies of the orientation of deformation lamellae in the Ajibik quartzite and other rocks. Fairbairn found that there was a stronger preferred orientation of the lamellae in the Ajibik quartzite than of the [0001]-axes, and inferred from this that the preferred orientation in the rock resulted from movements on the lamellae. The lamellae are inclined at variable angles to {0001}, but there is a strong maximum between 7° and 36°. The lack of a fixed crystallographic orientation for the lamellae obliged him to assume only a fixed glide-line [m:r], lying in the base, with variable glide-planes containing this line. Ingerson and Tuttle (1945) also demonstrated a conspicuous lack of crystallographic control in the orientation of the lamellae in the Ajibik quartzite, and their measurements show an even greater range of crystallographic orientation. These investigators also examined data for many grains with two or three sets of lamellae and concluded that these lamellae were not parallel to rational crystallographic planes. The angles between the lamellae and {0001} in the grains, however, did appear to be closely controlled by the orientation of individual grains in the fabric of the rock. Ingerson and Tuttle concluded that the lamellae are deformation-planes "such as would form in homogeneous material"; and that such crystallographic control as exists is only apparent, being actually dependent on the orientation of the grains with respect to the fabric axes. The poles of the lamellae in the Ajibik quartzite (as in many other rocks with quartz grains containing lamellae) form two strong maxima lying in the *ac* plane of the fabric and symmetrically oriented approximately 45° from the foliation. Ingerson and Tuttle claimed that in this and other specimens examined by them the pole of the lamellae in any grain lies between the [0001]-axis in the grain and the *a* fabric axis of the specimen. They suggested that this relationship might be used to locate the fabric axes in a rock, such as a massive quartzite, in which there is no other means of determining the fabric axes.

In his analysis of the microscopic structures in the Baraboo Quartzite, Riley (1947) found that deformation lamellae were extensively developed in rocks which showed very weak preferred orientations of the quartz grains. Applying the method suggested by Ingerson and Tuttle for locating fabric axes from the patterns of lamellae, Riley showed that the fabric axes thus obtained do not coincide with the fabric axes determined from megascopic data. He therefore inferred that, if the fabric axes determined from deformation lamellae are, in fact, related to the megascopic fabric axes in other areas, as Ingerson and Tuttle reported, the lamellae in the Baraboo Quartzite must date from a phase of deformation different from that which produced the foliation and folds.

Savul (1948) measured the orientation of deformation lamellae and optic

axes in different parts of individual deformed grains of quartz in a number of granitic rocks and quartz porphyries, selecting particularly grains with several sets of lamellae. On the basis of the angle between the poles of the lamellae and [0001], he distinguished 3 groups of lamellae; in one group the angles are between 8° and 36°, in the second, between 53° and 59°, and in the third, between 72° and 75°. These do not correspond to any of the common crystal faces in quartz. Savul superimposed the stereograms for measured quartz grains on a projection of a quartz crystal showing all the possible forms. This was also done for a number of quartz grains with several sets of lamellae and a few well-developed crystal faces. The lamellae, according to Savul, coincide most closely with the forms $\{10\bar{1}2\}$, $\{\bar{1}216\}$, $\{3\bar{1}22\}$, and $\{10\bar{1}4\}$ and he considered them to be produced by translation-gliding on these planes.

In the majority of cases in which deformation lamellae have been investigated in quartz tectonites, they have been interpreted as visible traces of the deformations which oriented the grains in the rocks. But Turner (1948) drew attention to incompatibility between the patterns of preferred orientation of lamellae and of other elements in the fabric in many of the examples described in the literature. Continuity of lamellae across grain boundaries (Mackie, 1947) implies that the grains must have had their present orientation in the rocks before the formation of the lamellae. Turner therefore suggested that the lamellae probably represent, in most cases, late-stage structures unrelated to the preferred orientation of the lattices of the grains. The lamellae are commonly oriented so that their poles define two strong maxima between 60° and 90° apart, a relationship which would indicate that the formation of the lamellae "is favored by simple compression (as in a vise) with consequent relatively slight differential movement upon surfaces of maximum resolved shear stress so induced" (Turner, 1948, p. 567).

Weiss (1954) has described the orientation of deformation lamellae and the somewhat similar deformation bands (Riley, 1947) in the quartzitic rocks of the Barstow area in Southern California. Weiss also considered that the lamellae were of rather late origin, and unrelated to the movements which produced the preferred orientation of the grains in which they occur.

## NEW DATA

*Introduction.*—Several authors have stated that deformation lamellae are of rather uncommon occurrence in rocks, but we have found that they are extensively developed in quartzites in metamorphic terranes in north-west Scotland (Christie, 1956) and Southern California. In the present paper the results of detailed microscopic analyses of four quartzite specimens are presented and an interpretation of the lamellar structures is proposed. Three of the specimens (I, II, III) are crystalline quartzites from the Orocopia Schists in the north-west corner of the Orocopia Mountains in Southern California, and the fourth (IV) is a deformed Cambrian quartzite collected a few feet below the outcrop of the Moine Thrust near the Stack of Glencoul, in the Assynt district of Scotland. The grains in all four specimens contain abundant deformation lamellae and other evidence of post-crystalline strain, such as undulose extinction.

*Nature of the structures examined.*—The planar structures whose orienta-

tions are described below may all be referred to as deformation lamellae. They are narrow, sub-planar structures which occupy a part or the complete area of a grain and generally are only found with one orientation in any grain. They are not all structurally similar when seen under the highest magnifications available: some consist entirely of minute brownish inclusions concentrated in planar zones; others cannot be resolved into individual inclusions and apparently have a different refractive index from the host grain; still others, intermediate between these two types, show a slight difference of refractive index and yet appear to consist in part of planes of inclusions (see Ingerson and Tuttle, 1945). Although the lamellae are gently undulating, they may all be measured on the U-stage, like cleavages and twin-lamellae in calcite and dolomite, by tilting them until they are parallel to the axis of the microscope. The maximum error involved is probably of the order of $\pm 2°$.

The above types of lamellae conform to the numerous descriptions in the literature, but it was noted that many of the lamellae have an extinction position which is slightly but markedly different from that of the neighboring part of the host grain. These lamellae are very conspicuous between crossed Nicols, particularly when the grain is close to the extinction position. The differences in the extinction position between lamellae and host grain are generally less than 3° (measured under high magnification on a flat stage) but an angle of 5° was noted in one grain.

Marked differences of optical orientation have been noted in the lamellar structures known as deformation bands (Riley, 1947; Weiss, 1954), but these are generally easy to distinguish from the more common deformation lamellae on the basis of the following criteria (Weiss, 1954):

1) Deformation bands are considerably broader than lamellae.
2) They are bounded by fractures which are not distinctly planar and whose orientation cannot be accurately measured with the U-stage; only the trend of the bands in the section can be measured.

The lamellae are generally, but not invariably, found in grains with marked undulose extinction in zones sub-parallel to the [0001]-axis, either with *continuous* or *discontinuous* change of extinction position ("plastic" and "ruptural" deformation of Hietanen, 1938).

Weiss (1954) found that the traces of deformation bands and deformation lamellae, when they occur in the same grain, are parallel, and he suggested that the two types of structure have the same genetic significance, the bands merely representing a more advanced stage of deformation than the lamellae. The present writers' discovery of small divergences of optic orientation between the material in deformation lamellae and in the parent grain is another point of similarity which supports Weiss' suggestion. The lamellae in which such differences of optic orientation may be observed do appear to be slightly broader than those in which differences are not detectable, but the lamellae are nevertheless well-defined and their orientation can be measured with the same accuracy as that of narrower lamellae. In a few of the grains examined the lamellae broaden into a wedge near the margin of the grain and differences of optic orientation are most readily seen in these wedges. No true deformation bands are present in the rocks.

*Description of specimen 1.*—This specimen (fig. 1a) is a folded quartzite

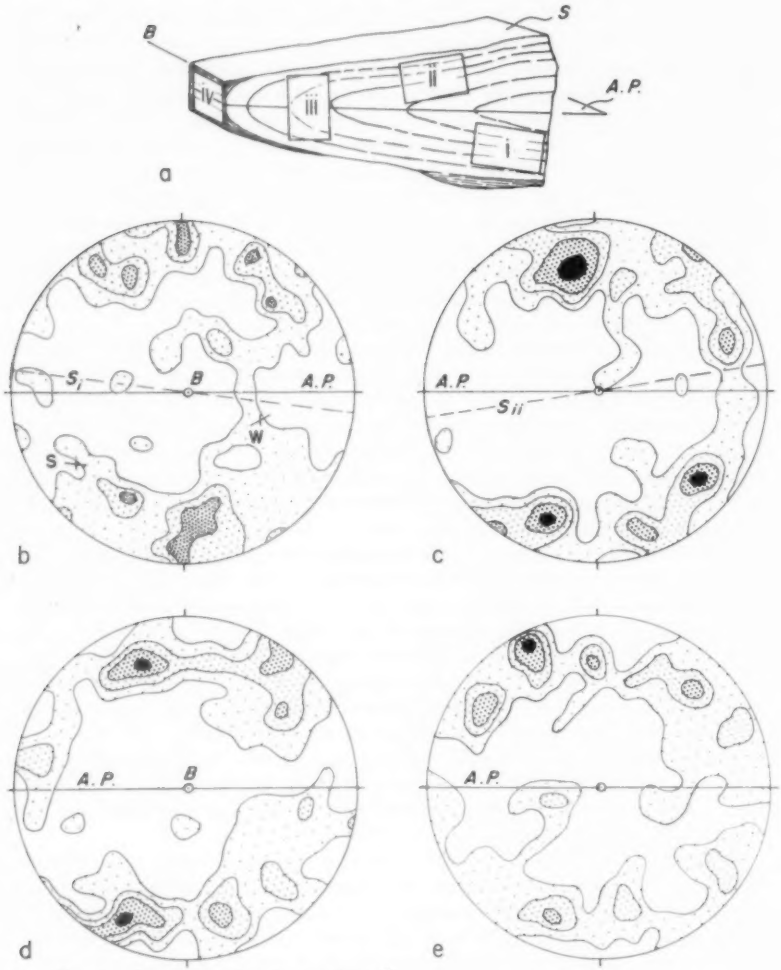



Fig. 1. Orientation data for specimen I.

a. Sketch of specimen I, showing the orientation and position of the thin-sections i, ii, iii and iv. *S* is the bedding-foliation, *B* is the fold-axis and *A.P.* is the axial plane of the fold.

b. [0001]-axes of 206 quartz grains in section i. Countours: 3, 2, 1% per 1% area.

c. [0001]-axes of 203 quartz grains in section ii. Contours: 5, 3, 2, 1% per 1% area.

d. [0001]-axes of 200 quartz grains in section iii. Contours: 5, 3, 2, 1% per 1% area.

e. [0001]-axes of 208 quartz grains in section iv. Contours: 5, 3, 2, 1% per 1% area.

Diagrams from all four sections have the same orientation, shown by south (S) and west (W) directions in diagram b.

with well-defined foliation from the Orocopia Schists. It consists essentially of quartz (95 percent) with minor amounts of calcite, plagioclase, garnet, chlorite, muscovite, biotite and opaque ore. The impurities are concentrated in thin layers which probably represent a relic bedding-structure. Under the microscope the quartz shows granoblastic texture with little or no dimensional orientation; the micaceous minerals are oriented with their cleavages subparallel to the axial plane of the fold. All the minerals, but especially quartz, calcite and chlorite, show evidence of post-crystalline strain.

In order to test the homogeneity of the preferred orientation of quartz in

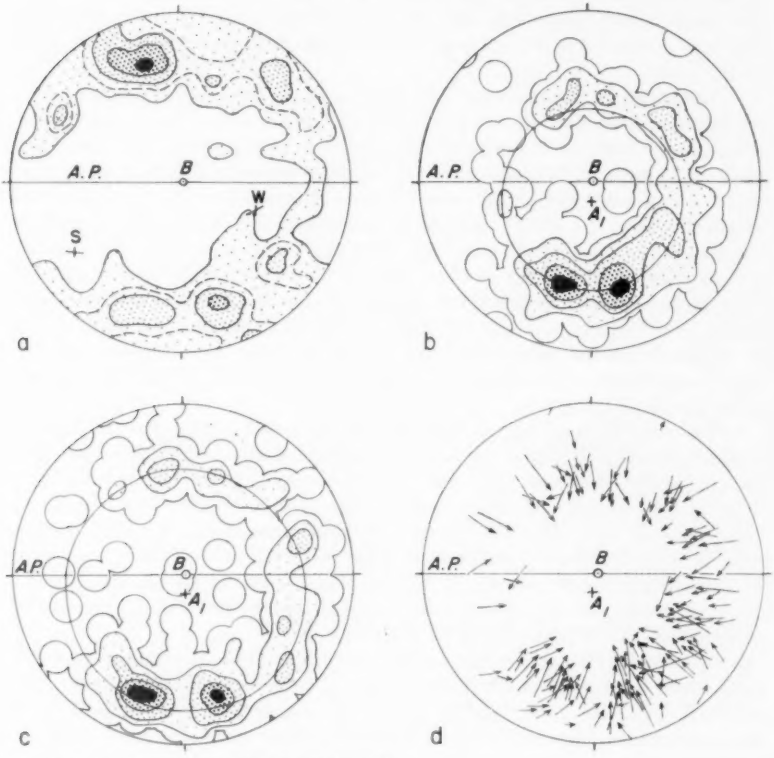



Fig. 2. Orientation data for specimen I.

a. Composite diagram: [0001]-axes of 817 quartz grains from the four sections. Contours: 4, 3, 2, 1½, 1% per 1% area.

b. Poles of deformation lamellae in 195 grains (195 sets of lamellae). Contours: 8, 5, 3, 1½, 1% per 1% area.

c. [0001]-axes of the same 195 grains containing deformation lamellae. Contours: 7, 5, 3, 1½, ½% per 1% area.

d. [0001]-axes (end of arrow) and poles of deformation lamellae (Point of arrow) in a representative number of grains from sections i-iv. $A_1$ is the axis of the small circle defined by the poles of lamellae and [0001]-axes of grains with lamellae. All four diagrams have the same orientation, shown by south (S) and west (W) directions in diagram a.

the specimen, analyses were carried out in four sections (fig. 1a). Three were cut parallel to the *ac* plane of the fold: one from each of the limbs (i, ii) and one from the hinge (iii). A further section was examined from a plan parallel to the fold axis (B) and approximately normal to the axial plane. The diagrams showing the orientation of [0001]-axes of quartz in the four sections are shown in figures 1b, 1c, 1d, 1e. The patterns of preferred orientation for all four sections are similar, indicating that the orientation of quartz is essentially homogeneous throughout the fold. The pattern of preferred orientation consists of a small-circle girdle, the axis of which is approximately normal to the axial plane of the fold (fig. 2a). The preferred orientation of micaceous minerals is also essentially homogeneous throughout the fold (diagrams not shown); the cleavage directions of the micas and chlorite are sub-parallel to the axial plane of the fold.

Deformation lamellae were measured in all the grains in which they were present and accessible for measurement with the U-stage. Each grain represented in the patterns of preferred orientation of quartz was carefully searched with the U-stage for lamellae, and it is therefore probable that in all the grains examined in each section all lamellae were measured except those inclined at low angles to the section and whose poles lie in the central "blind spot" in the diagrams. The partial diagrams showing the orientation of deformation lamellae in the individual sections show a strong similarity, indicating that there is homogeneity of this fabric element also throughout the specimen. The preferred orientation of the lamellae is such that the pattern obtained from the *ac* sections does not differ greatly from that obtained from the fold axis section; hence there must be few lamellae whose poles fall within the "blind-spots" in the diagrams.

The pattern of preferred orientation of the poles of lamellae (fig. 2b) is a small-circle girdle (about the axis $A_1$) containing two closely-spaced maxima and a number of sub-maxima. The poles of the lamellae therefore define (by their orientation) a cone with vertical axis $A_1$ and a semi-vertex angle of approximately 45°. The [0001]-axes of grains containing the deformation lamellae (fig. 2c) show a similar pattern of preferred orientation. They define a small-circle about the same axis $A_1$, but with a radius of approximately 60°; the small-circle contains maxima and sub-maxima which correspond closely with those in the pattern of lamellae poles (fig. 2b).

A striking feature of the grains with deformation lamellae is that the great-circle passing through [0001] and the pole of the lamellae generally passes through (or close to) the axis $A_1$ of the small-circle girdle. This is demonstrated in figure 2d, which shows the orientation of [0001] and the pole of lamellae in a representative number of the grains from all four sections.

*Description of specimen II.*—This specimen is a rather impure quartzite which is folded in a similar fashion to specimen I. The impurities comprise approximately 15 percent of the rock and include garnet, biotite, chlorite, muscovite and zircon. The garnets, zircons, and to some extent the micaceous minerals, are concentrated in layers defining the foliation, but the micas and chlorite are also disseminated throughout the purer layers of quartz. The micaceous minerals tend to lie parallel to the axial plane of the fold rather than the foliation. This tendency is most clearly observed in the hinge of the fold.

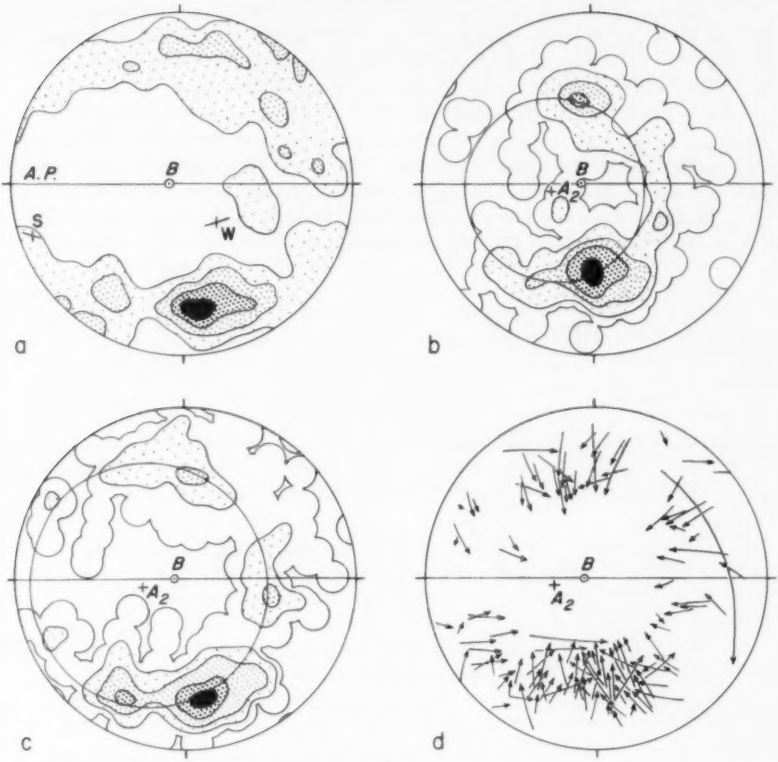



Fig. 3. Orientation data for specimen II.

a. [0001]-axes of 500 quartz grains, measured in three sections. Contours: 4, 3, 2, 1% per 1% area.

b. Poles of deformation lamellae in 207 grains (209 sets of lamellae). Contours: 9, 6, 4, 2, ½% per 1% area.

c. [0001]-axes of the same 207 grains containing deformation lamellae. Contours: 9, 6, 3, 1½, ½% per 1% area.

d. Poles of deformation lamellae (point of arrow) and [0001]-axes (end of arrow) in a representative number of grains from each section. *B* is the fold axis and *A.P.* is the axial plane of the fold. $A_2$ is the axis of the small circle defined by the poles of lamellae and [0001]-axes in grains containing lamellae. All four diagrams have the same orientation, shown by south (S) and west (W) directions in diagram a.

Analyses were made of the orientation of [0001]-axes and deformation lamellae in quartz in two *ac* sections of the fold, one from the hinge and one from a limb, and a *bc* section from one of the limbs.

The orientation of [0001]-axes of quartz is homogeneous throughout the fold, as in specimen I. The preferred orientation of [0001]-axes is shown in the composite diagram (fig. 3a). The pattern consists of a diffuse girdle normal to the fold-axis B, incomplete near the axial plane of the fold, and cleft at the pole of the axial plane; the girdle contains maxima inclined at high angles to the axial plane.

The poles of deformation lamellae (fig. 3b) in the quartz grains define a portion of a small-circle girdle which contains two maxima of unequal strength. The axis of the small-circle is $A_2$ and the radius is approximately 44°. The [0001]-axes of grains containing these deformation lamellae (fig. 3c) are oriented in a well-defined small-circle about $A_2$ and the small-circle contains a number of maxima which correspond more or less in orientation to those in the pattern of poles of lamellae (fig. 3b). The radius of the small-circle defined by [0001]-axes is approximately 60°. The great-circles containing poles of deformation lamellae and [0001]-axes in individual grains (fig. 3d) generally pass through the axis $A_2$, as in specimen I.

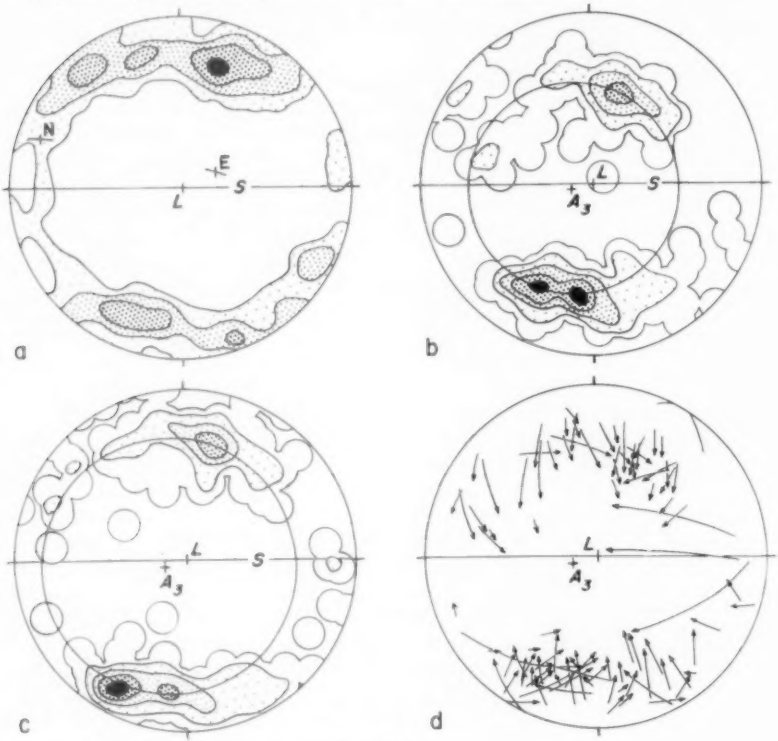



Fig. 4. Orientation data for specimen III.

a. [0001]-axes of 742 quartz grains, measured in two sections. Contours: 4, 3, 2, 1% per 1% area.

b. Poles of deformation lamellae in 144 grains (148 sets of lamellae). Contours: 8, 6, 4, 2, ⅔% per 1% area.

c. [0001]-axes of the same 144 grains containing deformation lamellae. Contours: 8, 6, 4, 2, ⅔% per 1% area.

d. Poles of deformation lamellae (point of arrow) and [0001]-axes (end of arrow) in a representative number of grains from each section. *S* is the foliation and *L* the lineation. $A_3$ is the axis of the small circle defined by the poles of lamellae and [0001]-axes in grains containing lamellae. All four diagrams have the same orientation, shown by north (N) and east (E) directions in diagram a.

*Description of specimen III.*—This specimen is a quartzite containing as impurities less than 10 percent of garnet, chlorite, biotite and opaque ore. It has a well-developed planar foliation (S), defined by layers of the above minerals, and a very weak lineation (L = B). Many of the quartz grains are slightly flattened in the plane of the foliation and there is some elongation parallel to the lineation.

The analysis was carried out on two thin sections cut normal to the foliation, one parallel and the other perpendicular to the lineation (*bc* and *ac* respectively). Approximately 350 grains were examined in the *ac* section and 390 in the *bc* section. Partial diagrams of 200 grains each from the two sections are almost identical, indicating a very high degree of homogeneity in the field of this specimen.

The pattern of preferred orientation of [0001]-axes (fig. 4a) consists of a girdle more or less normal to the lineation; the girdle is cleft near the pole of the foliation and the main maxima are inclined to the foliation at angles greater than 45°. The poles of the deformation lamellae (fig. 4b) define two maxima of unequal strength and tend to spread into a small-circle girdle about the axis $A_3$. The angular radius of the small-circle is approximately 52°. The preferred orientation of [0001]-axes of grains with lamellae is again similar to that of the lamellae poles, the radius of the small-circle being approximately 64°.

*Description of specimen IV.*—This specimen is a pure quartzite with a crude planar foliation and no lineation. The rock is a deformed sedimentary quartzite in which the original clastic grains are recognizable. The grains are considerably flattened in the plane of the foliation and show intense undulose extinction and moderate marginal granulation; zones of crushed quartz in the specimen are inclined to the foliation, which is defined only by the dimensional orientation of the large quartz grains.

The analysis was carried out on three mutually perpendicular sections, one of which was parallel to the foliation, and approximately 200 grains were measured in each section. The [0001]-axes of the quartz grains show a weak preferred orientation (fig. 5a). The pattern consists essentially of a diffuse small-circle girdle about the pole of the foliation, but there is also a strong maximum lying close to the foliation. This maximum is present in the partial diagrams from each of the three sections and must therefore be significant. The partial diagrams are otherwise similar only in that they all show the small-circle pattern referred to above; maxima in this girdle are not reproducible in the samples of 200 points. This is considered to reflect the weakness of the preferred orientation rather than lack of homogeneity of the fabric.

There is a very strong preferred orientation of deformation lamellae in the rock (fig. 5b). The pattern of preferred orientation consists of two strong maxima spreading into a small-circle girdle about the axis $A_4$, the radius of the small-circle being approximately 38°. The [0001]-axes of grains containing deformation lamellae also define a small-circle girdle about the axis $A_4$ with a radius of approximately 50°. Great-circles containing the pole of the lamellae and [0001] in individual grains (fig. 5d) commonly pass through the axis A, as in the other three specimens.

*Crystallographic orientation of lamellae.*—The histograms in figure 6

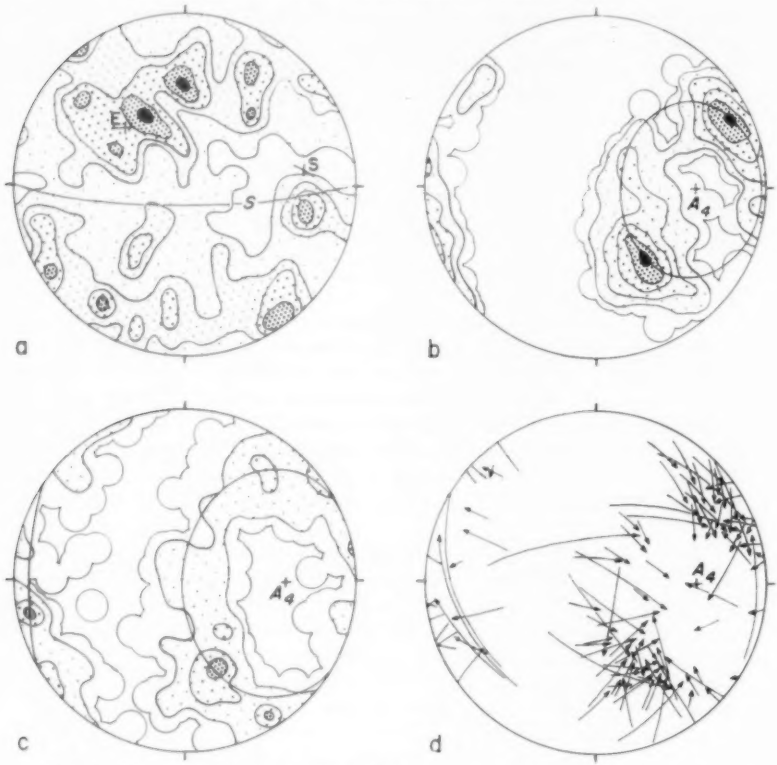



Fig. 5. Orientation data for specimen IV.

a. [0001]-axes of 607 quartz grains, measured in three sections. Contours: 2½, 2, 1½, 1, ½% per 1% area.

b. Poles of deformation lamellae in 204 grains (205 sets of lamellae). Contours: 10, 7½, 5, 2½, 1, ½% per 1% area.

c. [0001]-axes of the same 204 grains containing deformation lamellae. Contours: 4½, 3, 1½, ½% per 1% area.

d. Poles of deformation lamellae (point of arrow) and [0001]-axes (end of arrow) in a representative number of grains from each section. *S* is the foliation and $A_4$ is the axis of the small circle defined by the poles of lamellae and [0001]-axes in grains containing lamellae. All four diagrams have the same orientation, shown by east (E) and south (S) directions in diagram a.

show the angles between the pole of the deformation lamellae and the [0001]-axis ($C \wedge \perp L$) in the grains containing lamellae in each of the four specimens. The frequency distribution of these angles is similar in all the specimens: in the majority of the grains in each specimen the angle is between 0° and 40°, with a strong maximum in each case between 8° and 18°. The histograms are not dissimilar to those obtained by Fairbairn (1941) and Ingerson and Tuttle (1945) for the Ajibik quartzite, but there is a much greater variation in the ($C \wedge \perp L$) angle in the grains of the Ajibik rock. It is, of course, impossible to determine the orientation of the lamellae with respect to the hori-

zontal crystallographic axes of quartz in rocks of the type described above, but it is evident that the crystallographic orientation is highly restricted.

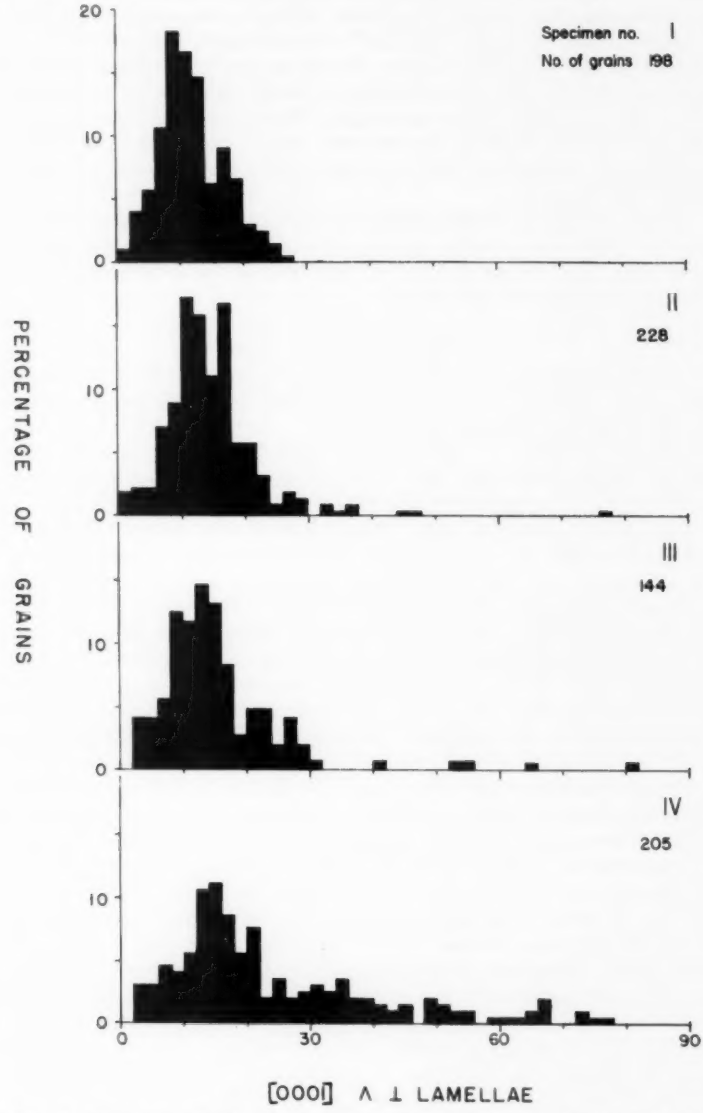



Fig. 6. Histograms showing the variation in the angle between [0001] and the pole of deformation lamellae in all the grains with lamellae in specimens I, II, III, IV.

*Summary.*—The orientation patterns of deformation lamellae in each of the four specimens described above are strikingly similar. In specimens II, III and IV, the poles of the lamellae from two distinct maxima 80° to 90° apart, indicating that the lamellae statistically define two surfaces; this is the type of pattern most commonly described in the literature (Turner, 1948). However, in all the specimens there is a more or less definite small-circle girdle containing the maxima. This distribution is also reflected by the [0001]-axes of the grains which contain the lamellae. The small-circle pattern of poles of lamellae was also noted by Riley (1947, fig. 8) in two samples of Baraboo quartzite. By comparison with the small-circle girdles, the maxima appear to be of minor importance.

The orientations of the great-circles containing the poles of the lamellae and the [0001]-axes in the quartz grains of all the specimens are similar. These great-circles (figs. 2d, 3d, 4d, 5d) intersect in, or close to, the axis A of the small-circle of lamellae-poles.

## INTERPRETATION OF DATA

*Discussion of fabrics.*—The folds in specimens I and II are "homogeneous" with respect to the orientation of quartz and mica (Sander, 1930, p. 260-262) and are therefore not unrollable.

When the grain orientation in such folds is symmetrologically related to the axial-plane of the folds they are commonly interpreted as shearfolds in which the folded surfaces are mechanically insignificant and the grain orientation has been produced by shear or flow on surfaces parallel to the axial-plane foliation (Sander, 1930). But a homogeneous grain fabric, particularly in the case of quartz, may also be produced by a strong imprint imposed on a fold of any origin (Sander, 1934, p. 44).

The patterns of preferred orientation of [0001] in specimens I, II and III are triclinic, if considered in detail, but there is a close approach, especially in specimens I and III (figs. 2a, 4a), to symmetry of a higher order—monoclinic or even orthorhombic. The closely appressed, almost isoclinal form of the folding (specimens I, II) precludes the possibility that the folds were formed purely by inhomogeneous slip on surfaces parallel to the axial-planes. Their development would require considerable flattening perpendicular to the axial planes to rotate the limbs almost into parallelism. Furthermore, there is good independent evidence, from the twinning of calcite in the rocks, that the folds have undergone strong compression normal to the axial planes late in their development.

The foliation in the quartzites is defined by zonal concentrations of accessory minerals in a framework of quartz grains and was therefore probably quite passive throughout the deformation. But the quartzites are thin and surrounded by comparatively incompetent greenschists, so that the boundaries were mechanically significant and evidently functioned as slip surfaces. We consider that the folds originated by flexural-slip, the slip being concentrated along the boundaries of the quartzite layers, and developed subsequently by rotation of the limbs resulting from compression along an axis normal to the axial planes. The quartz was undergoing recrystallization and reorientation throughout this deformational sequence. The main features of the patterns of

preferred orientation are considered to date from the late imprint of this deformation, when the strain had orthorhombic or higher symmetry, but the patterns preserve certain traces inherited from the earlier monoclinic movement.

The fabric of specimen IV differs from that of the other three specimens in that there is no evidence of complete crystallization of the quartz grains: the grains are relics of the clastic grains of the original sediment. The symmetry of the strain appears, however, to have been similar to that of the late paracrystalline imprint in the other specimens. Although recrystallization may have played a part in achieving the flattening of the rock, it was not sufficiently strong to obliterate the original texture. Plastic strain within the grains and intergranular shear appear to have been more important mechanisms of deformation.

*The origin of the deformation lamellae.*—Examination of the patterns of preferred orientation of the lamellae and of [0001]-axes in grains containing lamellae shows that in each rock these two elements show identical symmetry. But, as Turner has pointed out for other examples described in the literature (1948), the preferred orientation of the lamellae does not correspond in symmetry with that of the [0001]-axes in the rocks. This is particularly noticeable in the patterns for specimen IV; the pattern of preferred orientation of lamellae poles is almost perfectly orthorhombic and the three symmetry axes do not coincide with any of the fabric axes evident in the pattern of [0001]-axes (the 'flattening-foliation', S, and the axis of flattening, $\perp$ S). This lack of correspondence is also seen in the other specimens to a lesser extent; the symmetry of the pattern of lamellae poles is different from that of the [0001]-axes in each case.

From this we infer that the lamellae are secondary structures, dating from a late stage in the deformation of the rocks, and unrelated to the preferred orientation of the lattices of the quartz-grains (see Turner, 1948). They are not present in grains with all orientations in the general pattern of [0001]-axes. The patterns of preferred orientation of lamellae poles are not, however, so simple as those previously represented in the literature: although two maxima of poles, approximately 90° apart, are developed in the patterns for three of the specimens, a more constant feature of the patterns is the small-circle girdle about the axes designated A in the diagrams. The axial distribution of the lamellae poles in the specimens suggests that the lamellae have been produced by a simple compression (or tension) parallel to the axis A, as proposed by Turner (1948), but the surfaces of maximum shearing strain appear to be conic surfaces rather than two intersecting plane surfaces.

Before a further discussion of the origin of deformation lamellae is attempted, the nature of the structures must be considered. Several theories have been proposed to account for the orientation of deformation lamellae in the lattice of quartz: the lamellae have been considered as

a) Planes of twin-gliding (Judd, 1888) or translation-gliding parallel to rational crystallographic planes (Sander, 1930; Fischer, 1926; Schmidt, W., 1927; Savul, 1948)

b) Intragranular shear-surfaces with limited crystallographic control on

the orientation (Fairbairn, 1941; Turner, 1948; and to some extent Hietanen, 1938)

c) Intragranular shear surfaces with no direct crystallographic control on their orientation (Ingerson and Tuttle, 1945).

The histograms in figure 6 show that the orientation of the lamellae in the lattices of the grains is rather highly restricted, but certainly not to the extent that {01$\bar{1}$2} twin-lamellae are restricted in the lattice of calcite. A great variety of rational indices have been assigned to deformation lamellae by previous writers (Savul, 1948) but the following facts lead us to conclude that the lamellae are not parallel to rational planes in the quartz lattice:

a) The angle between the lamellae and the base {0001} varies *continuously* in any single specimen from 0° to 40°, and larger angles also occur.

b) There are slight variations in the position of the main maximum in histograms for different specimens.

The lamellae therefore are not parallel to rational crystallographic planes, but their orientation in the lattice is otherwise restricted, as Fairbairn (1941), Ingerson and Tuttle (1945) and Turner (1948) have maintained.

Ingerson and Tuttle (1945) accounted for the apparent crystallographic control over the orientation of deformation lamellae in the grains of the Ajibik Quartzite by demonstrating that the angle between the [0001]-axis and the pole of the lamellae in the grains varies systematically with the inclination of the [0001]-axes to the *B*-axis of the fabric; they concluded that the crystallographic control was only apparent and due to the fact that i) lamellae were cozonal with the B-axis and ii) the [0001]-axes were generally inclined at high angles to the B-axis. It is impossible, however, to account for the restricted crystallographic orientation of the lamellae in specimens I-IV in this way. Examination of diagrams 2d, 3d, 4d and 5d shows that the angle between [0001] and the pole of the lamellae in any grain is not controlled by the relationship between [0001] and any of the fabric axes defined by the foliation and fold-axis; nor is there any evidence from the patterns (figs. 2d, 3d, 4d, 5d) that the angle between the lamellae and [0001] varies systematically with the inclination of the [0001]-axes to the symmetry axes of the quartz [0001]-patterns. The restriction in the crystallographic orientation of the lamellae in these specimens is therefore not merely apparent (and controlled by the orientation of the grain lattices in the fabrics). It must be accounted for in some other fashion.

A striking and hitherto unexplained feature of the diagrams (figs. 2d, 3d, 4d, 5d), showing the relationship between the lamellae and the [0001]-axes in individual grains in the rocks, is the fact that, in each specimen, the great-circles containing the pole of the lamellae and [0001] pass through, or close to, the axes ($A_1$, $A_2$, $A_3$, $A_4$) of the small-circles defined by lamellae poles. This feature is so consistently evident (see also Riley 1947, fig. 8) that it must have some significance. The control of quartz orientation and, more particularly, the development of undulose extinction in zones sub-parallel to [0001] has commonly been attributed to limited bend-gliding on the surfaces represented by the deformation-lamellae (Johnsen and Becke in discussion of Schmidt, 1927; Hietanen, 1948). But the radial arrangement of the zones containing

[0001] and the poles of lamellae in the grains of quartz-tectonites would suggest that the reverse relationship may be correct: that the deformation lamellae originate as a result of movements parallel to the [0001]-axes of the grains. There is a mechanism, demonstrated by experimental studies, by which gliding movement on one set of surfaces may give rise to subsidiary movement along a plane inclined at steep angles to the glide-surface, namely, the formation of "kink-bands" (Turner, et al., 1954, p. 896-897). These kink-bands are slightly irregular zones, bounded by sub-parallel planes, in which the lattice of the mineral has a different orientation from that of the parent grain; they are inclined at high angles to the active slip-plane in the crystal and may or may not be parallel to a rational crystallographic plane.

In all the reported cases of minerals or metals in which kink-bands are considered to develop, they do so in response to gliding on a rational crystallographic glide-plane. There is no evidence, either from petrographic studies or from a consideration of the crystal structure of quartz (Griggs and Bell, 1938; Fairbairn, 1939) that such planes exist in quartz. However, there is good evidence that there is a linear weakness parallel to [0001]: the most obvious and common indication of post-crystalline deformation in quartz is the appearance of undulose extinction in zones sub-parallel to [0001]; when the strain is slight, the variation in extinction is continuous over a grain, but when the strain is more intense the grain is divided into distinct zones bounded by sharp surfaces of discontinuity. These boundaries are not distinctly planar and their orientation cannot be measured with the U-stage, indicating that they are not rational crystallographic planes but rather curved or irregular surfaces. The available data on the experimental deformation of quartz at high temperature and pressures (Griggs and Bell, 1938) show that quartz exhibits a strong tendency to break into needles parallel to the [0001]-axis. Examination of the crystal structure of quartz shows that none of the prism planes have the characters necessary to act as glide-planes, but it has been suggested (Griggs and Bell, 1938; Fairbairn, 1939) that imperfections in the crystal lattice, known as lineage structures (Buerger, 1934), which are known to exist parallel to [0001] in quartz, may be important in controlling the deformation. Griggs and Bell (1938, p. 1740) demonstrated that the needles parallel to [0001] are irregular and not bounded by prism planes.

Differential movement along [0001] of irregular blocks or rods elongate parallel to [0001] is part of the "Frontwendung" hypothesis of Hietanen (1938) and similar gliding movement is postulated by Weiss (1954). But in both of these hypotheses the deformation lamellae are considered to be the first indication of strain, representing planes of limited translation-gliding parallel to the basal pinacoid or rhombohedral planes. This is not in agreement with the actual crystallographic orientation of the lamellae. The following mechanism, which we consider to be the simplest and most rational means of accounting for the crystallographic orientation of the lamellae and their orientation in the fabric, is proposed:

Certain grains in the rocks have deformed by gliding parallel to the [0001]-axis, and the lamellae represent kink-bands inclined at high angles to the glide-surface and glide-direction. The type of gliding envisaged is not normal translation-gliding on a rational crystallographic plane; it may be

"*linear gliding*" involving differential movement of rods sub-parallel to the [0001]-axis (such as might be illustrated with a sheaf of pencils) or the gliding may take place on irregular surfaces of high resolved shear stress in the zone of [0001], involving differential movement of platy elements defined by these irregular surfaces.

In addition to explaining the restricted but non-rational orientation of the lamellae, this hypothesis accounts for most of the features observed in the patterns of preferred orientation of the fabric elements in the specimens. In a quartzite with random orientation of [0001]-axes, subjected to strong axial compression, gliding parallel to [0001] (or on non-rational surfaces in the zone [0001] ) will take place only in grains with an orientation such that there is high resolved shear stress suitable for gliding parallel to [0001]; the [0001]-axes of these grains define (in projection) a uniformly-covered small-circle girdle about the axis of compression. The stress in this instance is axial and the resulting strain will also be axial. However, if a rock with a strong preferred orientation of [0001]-axes is subjected to a similar axial compression, the grains oriented so that there is high resolved shear stress parallel to [0001] will again deform by gliding, but the pattern defined by the [0001]-axes of these grains will not be a complete and evenly-covered small-circle about the axis of compression: the pattern will be modified by the pre-existing preferred orientation. Although a small-circle may still be defined by the [0001]-axes of the deformed grains, it will contain maxima where the small-circle coincides with concentrations in the pre-existing pattern of [0001]-axes. (Although the stress in such a case is axial, the resulting strain, by this mechanism alone, would not be axial, but would have some lower symmetry controlled by the pre-existing anisotropy of the fabric.) An examination of figures 5a and 5c, for example, shows that the maxima of [0001]-axes in grains with deformation lamellae (fig. 5c) occur where the small-circle in this diagram crosses areas of high concentration in the pattern of preferred orientation of [0001] in the rock as a whole (fig. 5a). A similar relationship may be seen, though it is less evident, in specimens I (figs. 2a, 2c) and III (figs. 4a, 4c).

Kink-bands probably originate in response to some external influence which prevents unlimited deformation by gliding on a set of glide-planes with high resolved shear stress. In the case of deformed cylinders of calcite this influence is the constraint caused by clamping the ends of the cylinders, and thus preventing external rotation of the cylinder (Turner, et al., 1954). In the case of an aggregate of calcite or quartz grains, the constriction of any grain is probably imposed by the neighboring grains in the aggregate. This is illustrated for quartz in figure 7, assuming the mechanism postulated above, namely, gliding on non-rational surfaces of high resolved shear stress in the zone of [0001]. The development of kink-bands (represented by deformation lamellae and bands) in quartz may be explained as follows: When the aggregate of quartz grains (quartzite) is deformed by strong compression (fig. 7c) only grains which are suitably oriented for gliding parallel to [0001] (grains X and Y in fig. 7) will deform by gliding. As gliding proceeds these grains will become elongated and will rotate externally with respect to the axis of compression (figs. 7b, c). These processes are, however, impeded by the surrounding grains, which are not suitably oriented to deform by gliding. But

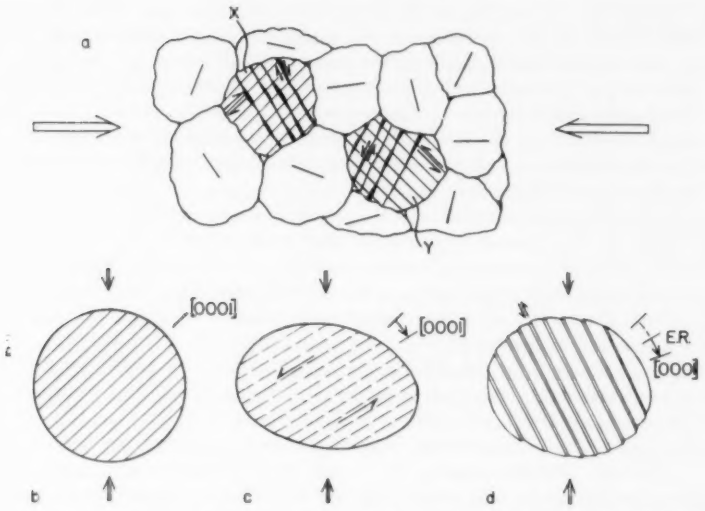



Fig. 7. Diagrammatic representation of the process of formation of deformation lamellae.

a. Aggregate of quartz grains, represented in two dimensions; uniform parallel lines in grains X and Y and short lines in other grains indicate the orientations of the [0001]-axes; irregular lines in grains X and Y represent deformation lamellae.

b,c,d. Change of shape of a grain of initially circular cross-section as a result of 1) shearing on surfaces parallel to [0001] and 2) kinking along planes inclined at a high angle to [0001]. The external rotation (E.R.) of the grain is shown qualitatively. Thick opposed arrows represent the direction of compression. For fuller explanation see text.

the production of kink-bands (fig. 7d) has the effect of correcting, at least to some extent, the change of shape induced by the gliding. Thus, if kink-bands are produced, gliding parallel to [0001] may take place with less change of shape than would be necessary in the absence of kinking.

It is probable that the undulose extinction and deformation lamellae are produced with a relatively small amount of post-crystalline strain, since in most of the specimens (I, II, III) there is no marginal granulation or marked elongation of grains. Moreover, the amount of strain in a grain containing kink-bands is reflected in the degree of external rotation of the kink-bands (that is, in differences of extinction position between the lamellae and host grain). These differences are very slight in deformation lamellae so that the strain must also be slight. It is significant that in highly deformed quartzites containing large amounts of granulated quartz and with very strong undulose extinction in the porphyroclastic relics, deformation lamellae are absent.[1] It is probable, in our opinion, that when deformation is so intense that differential movement and external rotation of grains occur, the restriction on individual grains is removed and deformation lamellae are no longer produced.

We do not wish to suggest that the mechanism postulated above is the

[1] This was noted by one of the writers (J.M.C.) in intensely deformed quartzites of the Assynt region.

only one which is operative in the deformation of quartz, or that it plays an important part in the development of preferred orientations in tectonites. There can be little doubt that solution and recrystallization are important in the reorienting of quartz and other mechanisms of plastic deformation may yet be demonstrated under conditions not yet attainable in the laboratory. However, the evidence described above leads us to conclude that it is significant in the initial stages of post-crystalline deformation of quartz and leads to the production of deformation lamellae.

A few deformation lamellae in the specimens may not have originated as kink-bands in the manner outlined above. Numerous cases are recorded, in specimens described herein and in the literature, of grains in which the lamellae are not inclined at high angles to the [0001]-axis. These lamellae may be shear-planes "induced" in a grain by the deformation of neighboring grains in the aggregate, either by gliding parallel to [0001] or by production of deformation lamellae in a neighboring grain. This view is supported by the existence of lamellae which pass more or less uninterrupted from one grain to another (Mackie, 1947, and present writers' observations). Moreover, some lamellae, particularly those which obviously consist of discrete inclusions and do not differ in refractive index from the host grain, may be relics of earlier fractures or deformation lamellae which have survived recrystallization of the grains (Hietanen, 1938).

*Summary of hypothesis.*—The deformation lamellae do not date from the deformation which produced the strong preferred orientations of the quartz grains in the rocks, but from a late penetrative deformation which has caused only slight strain and has had little or no effect on the preferred orientation. The grains have deformed by gliding on irregular surfaces in the zone of [0001] parallel to the direction [0001] and the lamellae are interpreted as kink-bands produced by this gliding. The strong preferred orientation of the lamellae is controlled by the preferred orientation of the grains in which such gliding has taken place, these grains being the ones which are oriented so that there is high resolved shear stress suitable for gliding parallel to [0001].

In the specimens described the [0001]-axes in grains with deformation lamellae define a small-circle girdle, representing a circular cone. The glide-surfaces are probably tangential to this cone and the poles of the induced kink-bands (deformation lamellae) also define a small-circle about the same axis. The orientation of the fabric elements suggests that the deformation was produced by a compression parallel to the axis of these small-circles. Two maxima are recognizable in some of the patterns of preferred orientation of lamellae, but the orientation and relative strengths of these maxima appear to be influenced by the existing preferred orientation in the rock before the production of the lamellae, rather than by the stress configuration during the deformation which produced the lamellae.

## APPLICATION OF THE HYPOTHESIS

*Significance of the proposed hypothesis.*—If the above hypothesis is reliable it should be of direct application to the *dynamic* interpretation of the microfabric of deformed rocks.

The axis of the small-circle girdles defined by poles of deformation lamel-

lae and [0001]-axes in grains with lamellae ('A' in figs. 2, 3, 4, 5) is the axis of maximum compression in the deformation which produced the lamellae. It is not the *a*-axis of the movement picture in this deformation, as claimed by Ingerson and Tuttle (1945). The patterns of preferred orientation of lamellae poles are not invariably small-circle girdles (Fairbairn, 1941; Ingerson and Tuttle, 1945; Riley, 1947) but in cases where this is not so, the axis A may be determined statistically as the intersection of the great-circles containing [0001] and the pole of lamellae in individual grains in the rock; in all the fabrics described in the literature, the pole of the lamellae in any grain is closer to the axis A than the [0001]-axis in the same grain (Ingerson and Tuttle, 1945). The orientation of the axis of compression (A) should be most easily determined in rocks showing little or no preferred orientation of the grain lattices, such as pure sedimentary quartzites and sandstones. But even in rocks with a strong preferred orientation of the quartz, such as schists and gneisses, it should be possible to identify the axis A using the above criteria.

By examining numerous specimens of quartzose rocks with deformation lamellae from an area it should be possible to obtain a reliable dynamic picture of the deformation which produced the lamellae. This should have more or less the same significance as the dynamic data obtained from studies of twinning in marble fabrics (e.g. Turner, 1953). For a terrane consisting of deformed crystalline rocks with strong preferred orientations of quartz, the inferred axes of compression will probably reflect only the last stage of the main deformation or some late imprint which has affected the rocks after the main phase of deformation and crystallization. However, for a terrane containing unrecrystallized quartz-rich sediments, the inferred axes of compression may reflect the main phase of deformation which has affected the rocks.

Certain precautions should be observed in the measurement and evaluation of data employed for such a dynamic interpretation:

1. Measurements should be made in two or three sections cut with different orientation from a specimen, to determine the real pattern of preferred orientation of lamellae.
2. The investigator should determine, if possible, that any lamellae found in sedimentary rocks are of post-diagenetic origin, and not relic structures which were present in the clastic grains of the sediment. The consistency of data, assessed by comparing successive samples of data from a single specimen, may indicate whether the lamellae are relics or have originated after the formation of the rock in which they are measured. Other criteria may also be found, such as the continuation of lamellae from a clastic grain in a quartzite into a peripheral region of secondary enlargement (Riley, 1947, p. 463).

*Application to data from the Baraboo Quartzite.*—The only regional study in which the orientation of deformation lamellae has been investigated is Riley's (1947) excellent study of the Baraboo Quartzite in Wisconsin. Riley made analyses of the orientation of [0001]-axes, deformation lamellae and other microstructures in the quartz grains of a large number of specimens from the Baraboo syncline. His detailed descriptions of the structures indicate that the deformation lamellae are identical with those described in the present work. The poles of deformation lamellae in most of the rocks define two strong

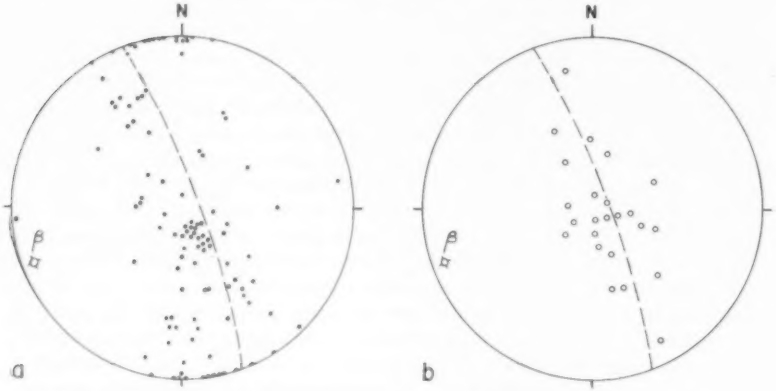



Fig. 8. Data from the Baraboo Quartzite (after Riley, 1947).

a. Poles of planar structures (bedding, shear-surfaces, etc.) in the Baraboo syncline. $\beta$ represents the regional fold-axis.

b. Compression axes ("*a*-axes" of Riley) deduced from the orientation of deformation lamellae in specimens of the Baraboo Quartzite. In both diagrams the primitive circle is horizontal with north (N) at the top.

maxima but in a number of the specimens they are oriented in a small-circle girdle, similar to those illustrated above (figs., 2b, 3b, 4b, 5b). Riley determined, according to the method suggested by Ingerson and Tuttle (1945), *a*-, *b*- and *c*- fabric axes from the patterns of preferred orientation of lamellae. These axes, however, showed no consistent relationship to the fabric axes determined in the field from megascopic structures such as foliation and folds, nor was there any obvious regional consistency in the orientations of *a*-, *b*- and *c*- axes determined from the patterns of deformation lamellae (Riley, 1947, fig. 14). Riley considered that the fabric axes thus determined might be either a) strain axes of a different deformation from that which produced the megascopic structures, or b) local strain axes of the deformation, independent of the movement along bedding surfaces.

If the patterns of preferred orientation of deformation lamellae are interpreted according to the theory proposed by the present writers, the axes designated "*a*-axes" by Riley (after Ingerson and Tuttle) are no longer to be considered as the directions of slip in planar deformations, but as *axes of compression* during the deformation which produced the lamellae. Figure 8b shows the orientation of these "*a*-axes" (compression axes) reproduced from Riley (1947, fig. 14). These axes lie in a vertical plane with NNW strike and many of the axes plunge steeply. Figure 8a is a $\pi$-diagram, also constructed from Riley's data, showing the orientation of the bedding surfaces in the area. The diagram shows a well-developed $\pi$-circle about the regional fold-axis which trends ENE. The shear-surfaces, axial plane cleavages and small folds agree in symmetry with this picture. It is evident that the axes of compression, determined from the deformation lamellae, lie in the *ac*-plane of the regional structure and also, therefore, in the deformation plane of the large-scale move-

ments. Thus the deformation lamellae may be related genetically to the movements which produced the large-scale folding of the Baraboo syncline. Although the movements may have been complex when considered in detail, as Riley has maintained, it is not impossible to incorporate the deformation lamellae and the folding in a single movement-picture.

ACKNOWLEDGMENTS

This paper reports the results of work carried out at Pomona College, Claremont, California, as part of a research project financed by the National Science Foundation; this support is gratefully acknowledged. We are indebted to Professor D. B. McIntyre for helpful discussion during the course of the study and the preparation of the manuscript. We are also very grateful to Professor F. J. Turner, who read the manuscript and made several very valuable suggestions; and to Professor D. T. Griggs, Professor G. Tunell and Dr. L. E. Weiss for helpful criticism and advice regarding the manuscript.

REFERENCES

Becke, F., 1892, Petrographische Studien am Tonalit der Riesenferner: Tschermak min. u. petr. Mitt., Band 13, p. 447.

Böhm, A., 1883, Über die Gesteine des Wechsels: Tschermak min. u. petr. Mitt., Band 5, p. 197-214.

Buerger, M. J., 1934, Lineage structure of crystals: Zeitschr. f. Krist. (A), Band 89, p. 195-220.

Fairbairn, H. W., 1939, Correlation of quartz deformation with its crystal structure: Am. Mineralogist, v. 24, p. 351-368.

Fairbairn, H. W., 1941, Deformation lamellae in quartz from the Ajibik formation, Michigan: Geol. Soc. America Bull., v. 52, p. 1265-1277.

Fischer, G., 1925, Mechanisch bedingte Streifungen an Quarz: Zentralblatt Min. Geol. Pal., Abt. A. no. 7, p. 210-213.

Griggs, D. T., and Bell, J. F., 1938, Experiments bearing on the orientation of quartz in deformed rocks: Geol. Soc. America Bull., v. 49, p. 1723-1746.

Hietanen, Anna, 1938, On the petrology of the Finnish quartzites: Comm. Geol. de Finlande Bull., no. 122, 118 p.

Ingerson, Earl, and Tuttle, O. F., 1945, Relations of lamellae and crystallography of quartz and fabric directions in some deformed rocks: Am. Geophys. Union Trans., v. 26, p. 95-105.

Judd, J. W., 1888, The development of a lamellar structure in quartz crystals: Mineralog. Mag., v. 36, p. 1-8.

Mackie, J. B., 1947, Petrofabric analyses of two quartz-piedmontite schists from northwest Otago: Roy. Soc. New Zealand Trans., v. 76, p. 362-368.

Mügge, O., 1896, Der Quarzporphyr der Bruchhauser Steine in Westfalen: Neues Jahrb. f. Mineralogie, p. 757-787.

Riley, N. A., Structural petrology of the Baraboo quartzite: Jour. Geology, v. 55, p. 453-475.

Sander, Bruno, 1911, Über Zusammenhänge zwischen Teilbewegung und Gefüge in Gesteinen: Tschermak min. u. petr. Mitt., Band 30, p. 281-314.

————, 1930, Gefugekünde der Gesteine: Vienna, Springer Verlag., v. 28, p. 37-50.

Savul, Mircea, 1948, Les orientations des plans de Böhm dans les cristaux de quartz: Inst. polytech. Jassy Bull., v. 3, p. 375-397.

Schmidt, W., 1927, Untersuchung über die Regelung des Quarzgefüges kristalliner Schiefer: Fortschr. der mineralogie, Band 11, p. 27-29.

Turner, F. J., 1948, Note on the tectonic significance of deformation lamellae in quartz and calcite: Am. Geophys. Union Trans., v. 29, p. 565-569.

————, 1953, Nature and dynamic interpretation of deformation lamellae in calcite of three marbles: Am. Jour. Sci., v. 251, p. 276-298.

Turner, F. J., Griggs, D. T., and Heard, H., 1954, Experimental deformation of calcite crystals: Geol. Soc. America Bull., v. 65, p. 883-934.

Weiss, L. E., 1954, A study of tectonic style: Calif. Univ., Dept. Geol. Sci., Pub., v. 30, p. 1-102.

[American Journal of Science, Vol. 257, June 1959, P. 408-430]

# A METAMORPHOSED DIFFERENTIATED SILL IN NORTHERN MICHIGAN*

RICHARD W. BAYLEY

U. S. Geological Survey, Menlo Park, California

ABSTRACT. An exceptionally thick differentiated metagabbro sill of middle Precambrian age has been found near Mansfield Location, Iron County, Michigan. This sill, known as the West Kiernan sill, intruded the sedimentary and volcanic rocks of the middle Precambrian Hemlock formation in late middle Precambrian time. The sill differentiated after emplacement, was subsequently folded, and altered by low-grade regional metamorphism. Differentiation resulted from mineral fractionation and gravity sorting during crystallization. Igneous textures and structures throughout the sill indicate tectonic quiescence during the time of crystallization. Rhythmic layering in part of the sill indicates that convection currents were probably active locally during crystallization.

The rocks of the sill have been divided into 5 zones, each representing petrographically and chemically distinct types. These zones are, from bottom to top, ultramafic, normal metagabbroic, transitional, granophyric, and metadiabasic (chilled).

Although the mineral composition of the sill was greatly changed during regional metamorphism, its chemical make-up is believed to chiefly reflect variations imposed during differentiation.

## INTRODUCTION

Mafic igneous rocks of the tholeiitic type representing two main periods of intrusion cut the Precambrian formations of northern Michigan. The most recent period, generally ascribed to Keweenawan time, is represented by a remarkable swarm of east-west trending, fresh, undeformed, diabase dikes, commonly showing negative magnetic polarization (Balsley, James, and Wier, 1949); whereas the older period, probably including much of late Animikie time, is represented by altered and deformed mafic dikes and sills showing normal magnetization and a less well-defined spatial arrangement. Sills of this older swarm are widespread in the area, and some of them are very large, but the West Kiernan sill of eastern Iron County (fig. 1) is the only one known to show vertical zoning resulting from differentiation.

In this report the geological relationships, petrography, and petrology of the West Kiernan sill are described briefly, and four new chemical analyses are presented.

The present investigation of the West Kiernan sill was carried out as a part of the general geologic mapping of the Lake Mary quadrangle by the writer (1956), and the Kiernan quadrangle by Gair and Wier (1956) under the auspices of the U. S. Geological Survey cooperating with the Geological Survey Division, Michigan Department of Conservation.

The writer wishes to extend his thanks to H. R. Cornwall, H. L. James, and C. D. Rinehart who made many helpful suggestions. I thank Lucille M. Kehl, Paul L. D. Elmore, and Katrine E. White of the U. S. Geological Survey for the chemical analyses cited. Photomicrographs used were prepared by personnel at the U. S. Geological Survey photographic laboratory, Washington, D. C.

* Publication authorized by Director, U. S. Geological Survey.

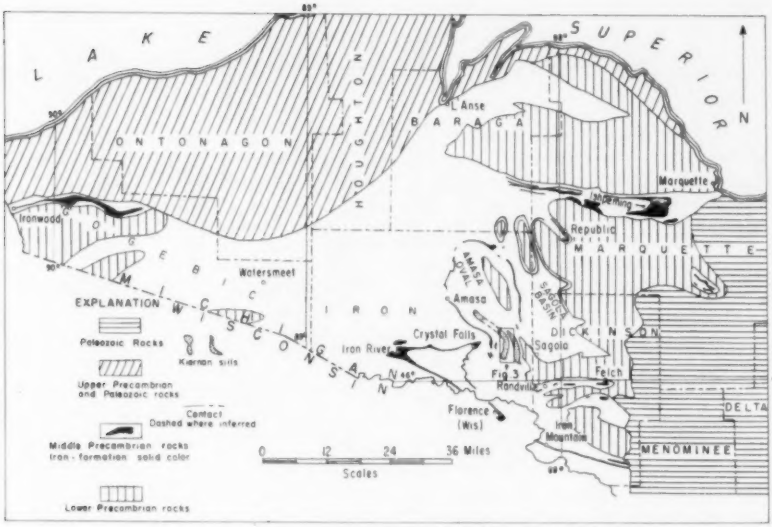



Fig. 1. Map of the western part of the Upper Peninsula of Michigan showing the location of the Kiernan sills with respect to the principal geologic features.

## GENERAL CONSIDERATIONS

There are two Kiernan sills, the East and the West, in eastern Iron County as shown on figure 1. The sills lie partly in the Kiernan quadrangle from which they take their name (Gair and Wier, 1956), and partly in the Lake Mary quadrangle (fig. 2). The East Kiernan sill is composed of normal metagabbro which shows no petrographic evidence of differentiation, and therefore will not be considered further here.

The West Kiernan sill forms a rugged, rocky upland along the east bank of the Michigamme River (fig. 2). The igneous character of the rocks and their general map outline were established by Clements (Clements and Smyth, 1899) who referred to the rocks as metadolerite. Micropegmatitic granite observed by Clements on the west flank of the metadolerite body near Mansfield Location, and interpreted by him to be a later intrusive, is now known to represent a distinct zone of granophyre genetically related to metadolerite. The total length of the sill probably exceeds 12 miles, but only the 7-mile segment shown on figure 2 has been mapped in detail. The average thickness is about 5,000 feet; the maximum, about 6,800 feet.

The location of the sill with respect to principal geologic features is shown on figure 1. The sill intruded the Hemlock formation (middle Precambrian) which is composed chiefly of ellipsoidal metabasalt (greenstone) and volcanic breccia, but contains beds of slate and iron-formation. The sill lies along the west limb of a major anticline which crests about 2 miles to the east. A normal west dip of about 70° is indicated by the bedded country rocks and by rhythmic layers within the sill. Ellipsoidal structures in metabasalt flows adjacent to

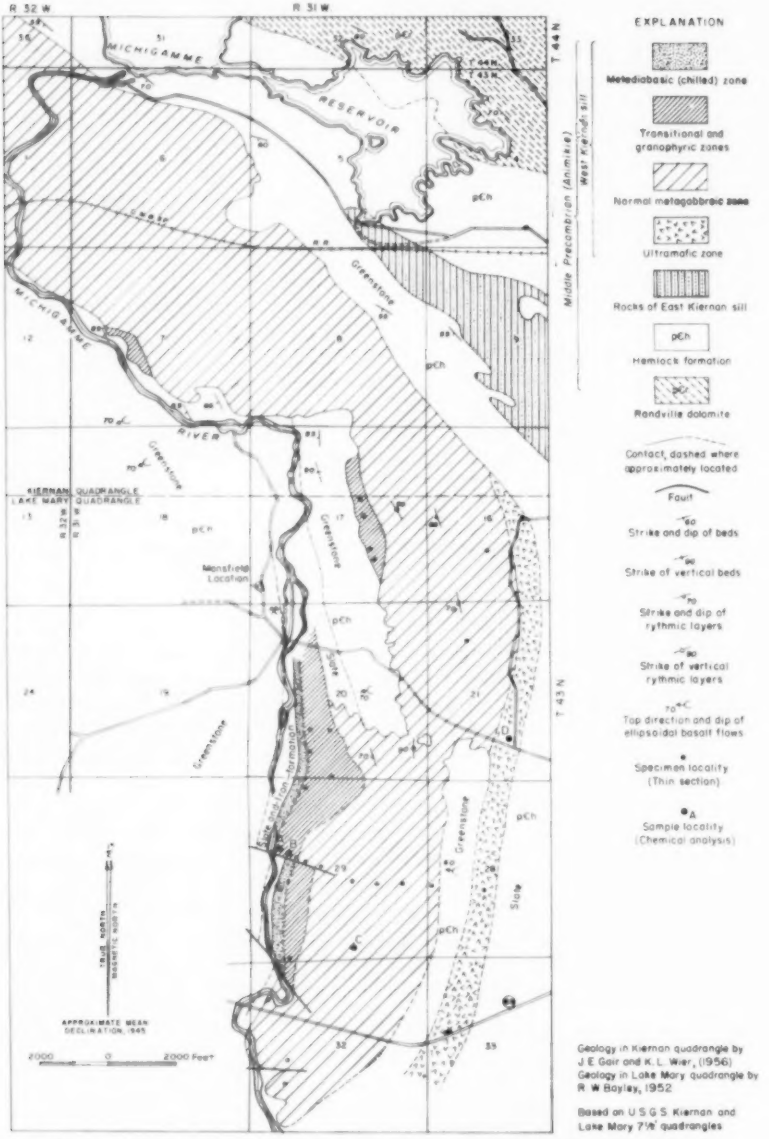



Fig. 2. Geologic map of part of the West Kiernan sill, Iron County, Michigan.

and enclosed within the sill show tops normal, facing west, and gravity concentrations of mafic minerals in the rhythmic layers are concentrated on the

east (bottom) side. The above field observations when coupled with petrographic and chemical considerations to be described below, lead inevitably to the conclusion that the intrusive is a sill; it differentiated while cooling in a horizontal position, and was subsequently folded to its present near vertical position.

The magma entered the Hemlock formation at shallow depth and raised the overlying rocks in the manner of a laccolith. Two sheet-like lobes of magma breached the country rock at slightly different stratigraphic levels, flowed toward one another, and ruptured across the intervening country rock, engulfing large segments of it where the two lobes met and commingled. The greenstone lying between the metagabbroic and ultramafic zones south of the east-west road from Mansfield Location (see fig. 2) is the same stratigraphic unit as the greenstone on the west (top) edge of the sill north of the above road. The displacement of the greenstone in the zone of rupture is about 3,000 feet. Structurally concordant sedimentary and volcanic rocks above (west of) the sill, particularly south of Mansfield Location, known by their metamorphic condition to have been present when the sill was intruded, were probably lifted 5,000 feet or more. At the bend of the Michigamme River, in the SW1/4 sec. 8, T. 43 N., R. 31 W., the upper contact of the sill jogs abruptly to the west, cutting discordantly across 5,000 feet of Hemlock formation, and from there continues northwest roughly parallel to the strike of the overlying rocks. It is evident from the above that the minimum thickness of the overlying rocks at the time of intrusion was in excess of 5,000 feet, but the total thickness cannot be determined.

For its total known length, the sill is in the low-grade area (chlorite and biotite zones) of regional metamorphism (James, 1955), and nearly all of the rocks examined conform mineralogically to the green schist facies as defined by Turner and Verhoogen (1951, p. 465). Contact metamorphism caused by the sill has been noted at many places along the top of the sill. Contacts with greenstone generally appear unaltered, but in a few places the greenstone within a few feet of the contact has been altered to a porphyroblastic actinolite hornfels. In view of the later regional metamorphism, (post-Animikie) the latter rock, where produced, was probably a hornblende or diopside hornfels. A slate and iron-formation member of the Hemlock is in contact with the sill south of Mansfield Location. Some of the slate is baked, porcellaneous, and some is chloritic and micaceous, and schistose. The latter type contains elongated pods of white mica, probably representing feldspar porphyroblasts formed during the contact metamorphism and destroyed during the later (post-Animikie) folding and regional metamorphism. The iron-formation, which is known from extensive diamond drilling in the area to be composed of chert and sideritic carbonate where unaltered, has been variously altered by contact metamorphism depending on the distance from the sill. Where slate forms a thin buffer zone between the sill and the iron-formation, porphyroblasts of stilpnomelane and magnetite are present in the chert and carbonate, but where the iron-formation is in direct contact with the sill or engulfed by it, grunerite, magnetite and quartz are the chief minerals present and the carbonate is wanting.

DESCRIPTION OF ROCK UNITS

The sill may be divided megascopically and microscopically into five distinct zones which, from the base upward, are (1) a basal ultramafic zone composed of serpentinized ultramafic rocks 600 to 1,200 feet thick; (2) a zone of diabasic and ophitic metagabbro 2,000 to 6,000 feet thick which commonly shows rhythmic layering; (3) a zone of iron-rich transitional rocks 200 to 1,000 feet thick, present locally; (4) a zone of granophyre 200 to 500 feet thick, present locally near the top of the sill; and (5) a zone of diabasic metagabbro 100 to 500 feet thick, probably representing the upper chilled margin of the sill, present locally. The zonation is shown on figure 2, but inadequate control requires combining the transitional and granophyric zones.

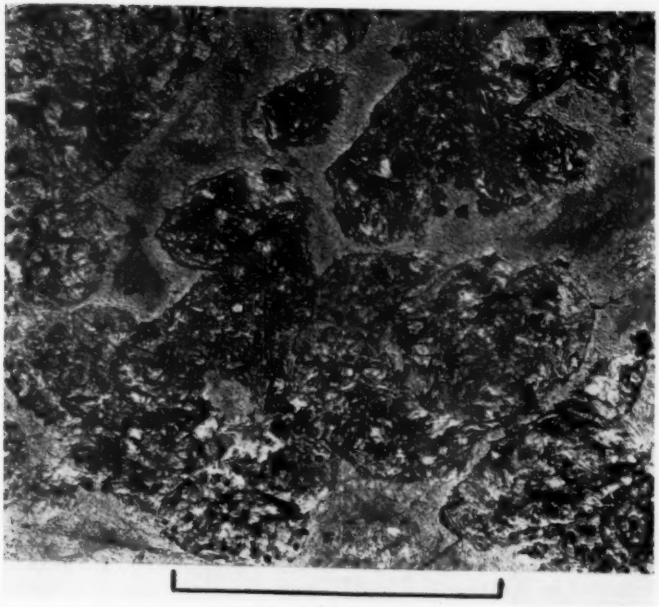
*Basal ultramafic zone.*—The rocks of the basal ultramafic zone are intermittently exposed in the area designated on figure 2. For over half of the length of the exposure belt the ultramafic rocks are separated from the main mass of the sill by extrusive greenstone of the Hemlock formation.

The rocks of this zone are serpentinized ultramafic rocks, probably peridotites and picrites. They are green to greenish-gray, medium-grained, and generally porphyritic. In hand specimens they characteristically show dull-green spots representing altered phenocrysts of pyroxene or olivine in a darker, greenish-gray, fine-grained matrix material. All specimens collected from this zone are moderately magnetic.

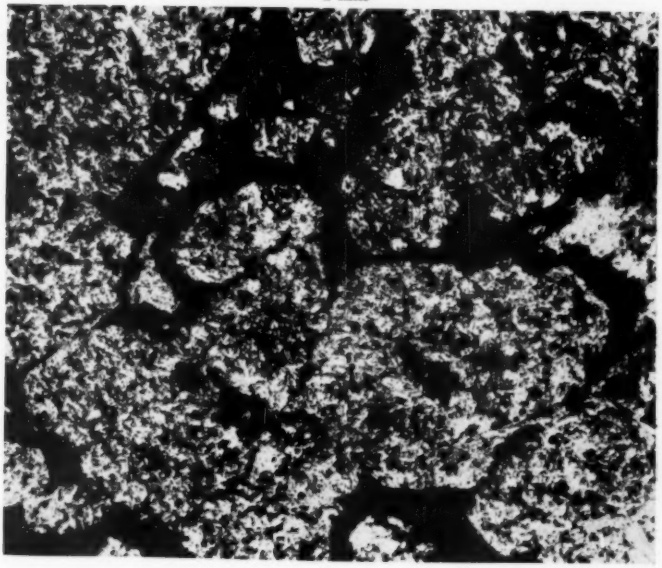
The minerals are serpentine, present in a variety of forms, tremolite, talc, magnetite, carbonate, chlorite, and actinolite. All are probably secondary minerals except perhaps some of the magnetite. Serpentine is present as pseudomorphs after pyroxene and possibly olivine phenocrysts and as the chief groundmass mineral. In some specimens the serpentine pseudomorphs are bastite containing ovoid inclusions of talc, tremolite, and magnetite which undoubtedly represent altered olivine crystals; in other specimens the serpentine phenocrysts are cryptocrystalline and contain tiny needles of tremolite which show preferred rectangular orientation. The form of the latter pseudomorphs is not distinctive enough to allow conclusive identification of the replaced minerals. Similar pseudomorphs were thought to represent altered olivine by Clements (Clements and Smyth, 1899) who described some metapicrite from northeastern Iron County, Michigan. In other specimens the pseudomorphic phenocrysts are composites of fibrous tremolite, talc, carbonate, and magnetite. These phenocrysts are anhedral, some of them oval shaped, and they lie in a very fine-grained serpentine and chlorite matrix. They are associated with phenocrysts composed essentially of tremolite that probably represent altered pyroxene; therefore the composite phenocrysts are presumed to represent altered olivine (fig. 3).

The ultramafic character of the basal sill rocks is shown by chemical analysis D, table 1. The rock analyzed is a medium-grained metaperidotite containing bastite phenocrysts in a groundmass of serpentine, tremolite, talc,

Fig. 3. Photomicrograph of serpentinized ultramafic rock from the basal ultramafic zone of the West Kiernan sill. × 50. A (above), ordinary light, B (below), crossed nicols. The groundmass is chiefly serpentine, and minor amounts of talc and chlorite. The phenocrysts are aggregates of tremolite, talc, carbonate, and magnetite, which may represent altered olivine.



1 mm



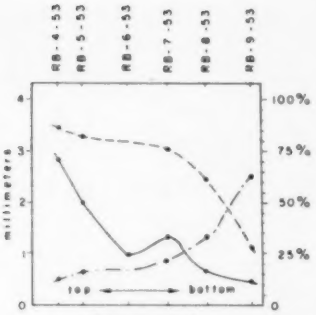
and magnetite. The analysis is similar to many analyses of picrite, peridotite, and kimberlites given by Washington (1917), and to average alkali peridotite and kimberlites of Nockolds (1954). Peak concentrations of the elements copper, cobalt, nickel, and chromium are particularly distinctive of this group of rocks. The normal and modal compositions of the metaperidotite of the West Kiernan sill are given on table 2. It has not been possible to determine how much of the mineral alteration of the rocks of the ultramafic zone was caused by regional metamorphism. Whatever their origin, the mineral assemblages now present are stable within the green schist facies of regional metamorphism. Assuming that a part of the alteration was premetamorphic (deuteric), alteration of the rocks during regional metamorphism has yielded similar mineral assemblages.

*Normal metagabbroic zone.*—The major rock type in the sill is normal metagabbro "dolerite" composed essentially of albite, actinolitic or tremolitic amphibole, and epidote. This rock makes up the bulk of the sill in the north where ultramafic and transitional zone rocks have not been recognized. In the south part of the sill, particularly south of Mansfield Location, the metagabbro forms a distinct zone, 3,000 to 4,000 feet thick, which grades down into the ultramafic zone and up into the transitional zone. The rocks of this zone show a distinct transition between the dark-colored, feldspar-lean ultramafic rocks and the light-colored, feldspar-rich transitional and granophyric rocks. Rhythmic layering similar to that found in other mafic stratiform sheets and described and illustrated by Wager and Deer (1939, p. 38-45) is a common structure in this zone. The layering is confined to the middle part of the zone, and massive metagabbro is more common near the margins. The individual layers are commonly less than 6 inches thick and each is stratified, probably by gravity sorting, with upward enrichment in feldspar and downward enrichment in mafic minerals. Primary igneous lineation caused by the alignment of elongate plagioclase crystals parallel to the top of the sill has been observed in a few places near the top of the normal metagabbroic zone, but it is not a conspicuous feature and may have been overlooked at other places. The alignment of crystals is not known to be associated with rhythmic layers, and therefore was possibly caused by fluxion.

Sparse pegmatite stringers composed chiefly of plagioclase, apatite, titaniferous magnetite, and blades of actinolite up to six inches long occur in a few places near the top of the zone.

Six specimens collected along an east-west traverse across the zone near the center of sec. 29, T. 43 N., R. 31 W., show an increase in the amount of original plagioclase from about 30 percent at approximately 2,000 feet above the base of the sill, to about 87 percent near the transitional zone at about 5,000 feet above the base of the sill (fig. 4).

In the same group of specimens, the plagioclase laths show a general increase in average length from about 0.4 mm in the lower part of the zone to 3.5 mm in the upper part of the zone (fig. 4). The maximum length of plagioclase crystals observed in the upper part of the zone is about 2 cm, but in general the crystals are smaller. Relict diabasic and ophtic textures are prevalent in these rocks and are megascopically distinct. The ophitic rocks are particular-




| | RB-4-53 | RB-5-53 | RB-7-53 | RB-8-53 | RB-9-53 |
|---|---|---|---|---|---|
| Specific gravity | 2.94 | 2.94 | 2.87 | 2.93 | 2.99 |
| Plagioclase (An 5-35) | 87.0 | 82.0 | 78.0 | 65.0 | 30.0 |
| Amphibole-chlorite-serpentine | 12.5 | 16.5 | 21.0 | 34.0 | 66.5 |
| Titanite-rutile-magnetite | 0.5 | 1.5 | 1.0 | 1.0 | 2.8 |
| | 100.0 | 100.0 | 100.0 | 100.0 | 99.3 |

Fig. 4. Modes of five specimens from the normal metagabbroic zone. On the graph, percentages of plagioclase and amphibole, and the length of the average plagioclase lath in each specimen are plotted against their location within the normal metagabbroic zone.

ly striking in appearance showing poikilitic amphibole individuals 1 to 2 inches in width.

The minerals of the normal metagabbro are chiefly secondary. They are albite and oligoclase ($An_5$ to $An_{35}$), actinolite, tremolite, chlorite, serpentine, epidote, clinozoisite, calcite, stilpnomelane, ilmenite, sphene, magnetite, quartz, and potassium feldspar. Albite and actinolite or tremolite make up about 90 percent of most of the rocks, but in some specimens other alteration minerals are dominant.

A few specimens from the lower part of the zone contain relatively unaltered labradorite ($An_{55}$), but in general the plagioclase has been replaced by aggregates of albite, clinozoisite of chlorite, and epidote (fig. 5). The original calcic character of the replaced plagioclase is indicated by the high anorthite content ($An_{72}$) of the normative plagioclase of analysis C, table 1 and table 2.

Granophyric intergrowths of quartz and potash feldspar occur sparingly in the rocks near the top of the metagabbroic zone, and increase in amount upward (westward) to become dominant in the granophyric zone. The granophyric intergrowths occur interstitially among the plagioclase and amphibole of the metagabbro as the last minerals to crystallize.

Amphibole is present in a variety of forms and colors, and is variously replaced by other minerals. It occurs chiefly as poikilitic individuals of large size in the ophitic rocks, and as anhedral fillings in the angular spaces between the plagioclase laths in the diabasic rocks. In general the amphibole is colorless and tremolitic in the lower and central rocks of the zone, and becomes pale

Fig. 5. Photomicrograph of diabasic metagabbro. × 50. Ordinary light. Showing saussuritized laths of palgioclase (clinozoisite-epidote-chlorite-albite), Sp; and actinolitic amphibole, A, pseudomorphic after pyroxene.

green and actinolitic toward the upper part of the zone, perhaps reflecting an increase in iron in the original pyroxene. Many characteristics of the original pyroxene are retained by the secondary amphibole; these include stubby prismatic growth, simple and polysynthetic twinning, and, in a few specimens, a well-developed basal parting. The secondary amphibole in some of the rocks is partly replaced by chlorite, serpentine, epidote, clinozoisite, and stilpnomelane.

Stilpnomelane replacement is particularly common in the upper part of the zone.

Ilmenite and magnetite occur as conspicuous accessories, comprising up to five percent of some rocks high in the zone. The ilmenite forms euhedral and anhedral skeletal crystals, poikilitic plates, and rods with rounded terminations. Most of these are altered to leucoxene or sphene or rutile and magnetite. Rocks containing ilmenite and magnetite as principal accessories are moderately magnetic.

A chemical analysis of a sample collected near the center of the sill is shown in column C, table 1. The norm and mode of the same sample appear on table 2. The rock analyzed is composed chiefly of the alteration minerals, clinozoisite, albite, tremolite, chlorite, and sphene. It has relict ophitic texture with 15 to 20 percent of nearly colorless poikilitic amphibole distributed in broad patches. The plagioclase laths are almost completely replaced by clinozoisite, albite, and minor amounts of chlorite. The secondary, hydrous mineral suite is indicated in the analysis by an excess of water; in other respects, the analysis compares most favorably with average analyses of gabbros and basalts (Daly, 1933, p. 17, and Nockolds, 1954, p. 1020-1021).

The rocks of the normal metagabbro zone are with few exceptions greenstones mineralogically commensurate with the green schist facies of regional metamorphism. The metabasalts of the host Hemlock formation are in equilibrium in the same facies for thousands of feet both above and below the sill (Bayley, 1956).

The characteristic mineral suite of the metagabbro includes albite-actinolite-epidote-chlorite, and other minor constituents. But in all of the rocks the plagioclase has not reached equilibrium, with the result that some saussuritized rocks contain remnants of plagioclase containing up to 35 percent anorthite. In one small area near the center of the sill the rocks contain relatively unaltered labradorite.

The probable sequence of mineral transformations during metamorphism is indicated below:

Labradorite — pyroxene — ilmenite — magnetite

(1) Labradorite — actinolite — sphene — magnetite

(2) Albite — actinolite — epidote — chlorite — sphene — magnetite

The fact that primary igneous textures prevail in the metagabbro indicates that the metamorphism was essentially low-grade hydrothermal. The pervasive character of the alteration of most of the rocks of the district indicates that heat and water were each available in sufficient quantities to transform thoroughly all amenable rocks to the final stable assemblage. The anomalous local occurrence of metagabbro which contains actinolite and original labradorite in the body of the sill, may indicate inequal distribution of water during metamorphism, probably caused by the low permeability of the gabbro and the great thickness of the sill. Probably with less water available than is needed to form assemblage (2), above, the pyroxene altered to amphibole preferentially and the labradorite remained unaltered.

*Transitional zone.*—The transitional zone includes a group of highly altered rocks that seem to be mineralogically and texturally transitional between the metagabbro and the granophyre. The rocks are generally dark brown and fine-grained; they are soft and weather into coarse angular rubble. Tiny needles of apatite and grains of dark quartz are the only minerals visible in a hand specimen. These distinctive rocks form a fairly persistent belt of varied thickness below the granophyric rocks in the upper part of the sill. The transitional rocks developed only where the sill is concordant with the overlying sediments. In the SE1/4 sec. 7, the SW1/4 sec. 8 and the NE1/4 and SE1/4 of sec. 20, T. 43 N., R. 31 W., where discordant contacts prevail, neither granophyre nor transition rocks were detected, nor are the transitional rocks known from the northwest part of the sill.

Most of the rocks of the transitional zone show vestiges of an original diabasic texture, but some of them show hypautomorphic-granular or xenomorphic-granular textures. The minerals present are the same as in the metagabbro below, with these exceptions: (1) ferric stilpnomelane is abundant, which accounts for brown color of the rocks; (2) apatite, a minor accessory mineral in the metagabbro below, is moderately abundant as long, thin, euhedral needles that cut all the major mineral boundaries; (3) altered ilmenite, and magnetite are present in amounts up to 5 percent; (4) the amphibole is darker colored, in green, blue-green, or brownish-green, and is often color mottled; (5) free quartz and micropegmatite are present in moderate amounts; and (6) euhedral blue-gray tourmaline is present in trace amounts.

Amphibole of similar description and a blue-green tourmaline have been described by Leighton (1954) from the transitional rocks of the northern Wisconsin gabbro-granophyre complex. Blue-green tourmaline has also been noted by Blackader (1956) in red-rock associated with the Logan sills of Ontario.

The amphibole, where present, occurs only as remnants, partly replaced by cryptocrystalline stilpnomelane and chlorite, and epidote. In some specimens the stilpnomelane also replaces the potassium feldspar fraction of the micropegmatite. Chlorite selectively replaces the cores of many plagioclase crystals, and in specimens with remnant diabasic texture it is aggregated with fine-grained albite in the slots once occupied by the original plagioclase. In other specimens the plagioclase has been partly replaced by granular aggregates of epidote and clinozoisite, and in such rocks, the plagioclase remnants the either albite or albite-oligoclase.

Chemical analysis B, table 1, represents a typical rock sample from the transitional zone. The norm and the mode of the anlyzed rock are given on table 2. Analysis B does not compare favorably with either Daly's or Nockold.' average igneous rock types, and is probably a hybrid. The chemical composition of the Kiernan transitional rock is closely approached in composition by a transitional rock from northern Wisconsin whose analysis is shown for comparison in column (2), table 3. The Wisconsin specimen is a metagabbro which according to Leighton (1954, p. 426) has been altered metasomatically by an intrusive granophyre.

It has been impossible to make any distinction between the effects of regional metamorphism and the effects of deuteric alteration in the transition-

al rocks. Because hydrous mineral alteration is a characteristic of gabbro-granophyre transitional rocks, and because such alteration may be extreme in rocks not affected by later regional metamorphism (Bastin, 1938), it is reasonably certain that at least part of the alteration of the transitional rocks of the Kiernan sill was deuteric. Assuming that such alteration did take place late in the cooling history of the sill, conditions suitable for hydrous alteration were re-established in the transitional rocks by later regional metamorphism and some of the alteration of these rocks undoubtedly dates from that period.

Figure 6 illustrates to what degree the three analyzed specimens of metamorphic rock from the Kiernan sill have approached equilibrium in the green schist facies, insofar as the mineralogy is directed by chemical composition. The three analyses plotted contain excess $Al_2O_3$; analyses B and C contain excess $SiO_2$ as well. Note that point D will approximately plot at D′ when com-

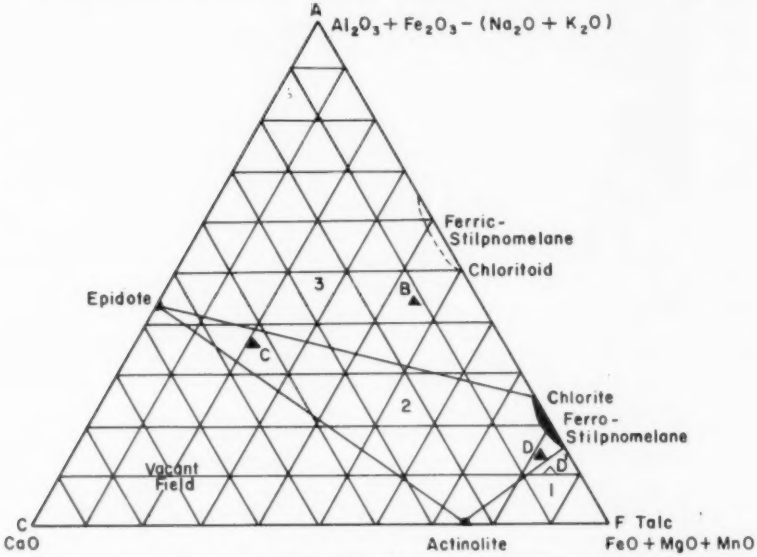



Fig. 6. Green schist facies, muscovite-chlorite subfacies; lettered points correspond with chemical analyses on table 1. ACF diagram slightly modified from figure 71, p. 471, of Turner and Verhoogen (1951).

pensation for surficial oxidation of the sample is made by arbitrarily reversing the quantities of ferrous and ferric iron. Point C, assemblage (2) in the diagram, contains the requisite mineralogy for that assemblage (mode C, table 2). The chief modal constituents of the rock represented by point B are ferric-stilpnomelane and chlorite (mode B, table 2). The predicted mineralogy of the specimens therefore almost perfectly coincides with the modes.

*Granophyric zone.*—Granophyre is present near the top of the sill in secs. 7, 17, 20, and 29, T. 43 N., R. 31 W., but the limits of the zone are not well established. The upper contact of the zone has not been observed, but an ap-

parent gradation between the rocks of the normal gabbroic, transitional, and granophyric zones has been observed in nearly continuous exposures in the NW¼ sec. 29, T. 43 N., R. 31 W. The granophyric zone is discontinuous along the strike and is presumed to be lenticular. It is confined chiefly to areas where the upper contact of the sill is nearly accordant with the overlying country rock. In the SW¼ sec. 8, and the NE¼ sec. 20, T. 43 N., R. 31 W., there are large scale discordances along the upper contact of this sill, and both transitional rocks and granophyre are absent.

The granophyre, as the name implies, is characterized by micrographic texture. It bears a close resemblance to granophyre described from other differentiated mafic sills. The appearance of the rock in thin section is most distinctive, but as Johannsen noted, it is easier to recognize than describe.

Two main varieties have been distinguished in the field, (1) dull greenish-gray, fine- to medium-grained granophyre showing small grains of red feldspar and dark blebs of glassy quartz in a schistose base material, and (2) buff to reddish-brown, medium-grained, granitic-appearing granophyre, either showing no mafic mineral, or showing abundant needle-shaped aggregates of magnetite or, if surficially altered, ocher-colored iron oxide.

The essential minerals of both varieties are albite or albite-oligoclase, potassium feldspar (microcline and orthoclase), and micropegmatite. Free quartz is an important constituent in some specimens. Secondary and accessory minerals are chlorite, sericite, biotite, epidote, calcite and a ferruginous carbonate, apatite, sphene, magnetite, and zircon.

The feldspar is generally clouded by fine-grained alteration minerals, chiefly chlorite and sericite, and an undetermined red dust, possibly hematite. Chlorite is the chief mineral of the schistose base which distinguishes one variety of granophyre.

In each variety the feldspars are similar, but relative quantities vary. The sodic plagioclase forms equidimensional to slightly elongate crystals commonly showing one large central zone, untwinned or twinned according to the albite and Carlsbad laws, and one outer zone or rim of outward radiating micropegmatite or microcline. The pericline twin of the microcline, in effect, gives the appearance of a radiating outer zone as does the micropegmatite. Many crystals show an irregular penetration of the microcline twinning into the central zone. Both orthoclase and microcline are abundant, but the relative amounts of these have not been determined. Both are intergrown with quartz and both form anhedral or subhedral crystals many of which show Carlsbad twinning (fig. 7).

Of the accessory and secondary minerals listed above, sphene, biotite, and apatite are virtually restricted to the first (schistose) variety of granophyre. The sphene forms skeletal pseudomorphs after ilmenite which are set in halos of fine-grained dirty-brown biotite. The apatite occurs as euhedral needles as in the transitional rocks.

Secondary calcite is pervasive. What appears to be a primary carbonate is the only important accessory mineral in the variety of granitic appearing granophyre which shows no mafic mineral in hand specimen. In all specimens examined the carbonate is partly replaced by hematite or limonite; it shows high indices relative to balsam, and therefore is presumed to be ferruginous.

Fig. 7. Photomicrograph of granophyre. × 65. Crossed nicols. Showing implication of microcline (m) and quartz (q); Carlsbad twins of microcline, and orthoclase (o). Cut by veinlet of carbonate (c).

The granophyre which contains needle-shaped aggregates of magnetite is, in other respects, similar to the granophyre containing the carbonate. Also, the magnetite shows the same paragenetic relationships as the carbonate, indicating that either the magnetite replaced the primary carbonate or that the magnetite

formed as a primary mineral under slightly different chemical conditions than those prevailing where the carbonate formed. Surficial oxidation has altered a considerable part of the magnetite in the exposed rocks to earthy iron oxide.

A chemical analysis of a chip sample representing the schistose variety of granophyre is given in column A, table 1; the norm and mode of the sample are given in table 2. Modes of other granophyre specimens are given in table 4.

The effects of regional metamorphism are generally indeterminate in the rocks of the granophyre zone. Some chloritization and sericitization, as noted above, may date from the regional metamorphism, but for the most part, the character of the rocks could be attributed to their primary igneous history.

*Metadiabasic (chilled) zone.*—A zone of fine- to medium-grained metadiabase 100 to 400 feet thick, probably representing the chilled upper selvage of the sill, is present on the west flank of the sill in secs. 20 and 29, T. 43 N., R. 31 W. (fig. 2). The metadiabase is generally concordant with the country rock, but at a few places in the SW¼ sec. 29, a few large elongate bodies of meta-iron-formation (grunerite-magnetite-quartz hornfels) are included within it.

Texturally and mineralogically the metadiabase resembles the rocks of the normal metagabbroic zone and the metabasalts of the Hemlock formation, but it is finer-grained than much of the former and coarser-grained than the latter.

The metamorphic mineral suite of the metadiabase consists of albite-actinolite-epidote-chlorite. There are no relict minerals, but the igneous (diabasic) texture of the rock has not been destroyed. The grunerite hornfels within the metadiabase presents an anomaly in the regional metamorphic picture, in that the hornfels represents at least garnet-zone metamorphism, whereas the grade for the area as reflected in the other rocks, including the metadiabase wherein the hornfels resides, indicate chlorite grade generally. It is clear that the hornfels was produced from chert-carbonate iron-formation by the heat of the basaltic magma which formed the diabase—it is interesting to observe that the hornfels thus formed does not reflect the later regional metamorphism which completely altered that diabase.

## DISCUSSION OF CHEMICAL DATA

The chemical analyses of four samples representing the four lower zones of the West Kiernan sill are shown in table 1. The locations where the samples were collected are shown on figure 2. On figure 8 the perecntage of the main oxides are plotted against depth within the sill, which, assuming that the sill crystallized from the bottom up, may roughly represent the percentage crystallized. Because of the irregular spacing of sample localities, the position of samples C and D have been projected into the section containing samples A and B. The variation of oxides as shown on the diagram, in general, show a reasonable correspondence with those determined for unmetamorphosed differentiated bodies (Wager and Deer, 1939, pl. 27; Walker and Poldervaart, 1949, p. 151-154; Cornwall, 1951, p. 156-158; and Hotz, 1953, p. 694). The variations particularly of the oxides $SiO_2$, CaO, $TiO_2$, MgO, $K_2O$, $Na_2O$, $P_2O_5$, and MnO conform well to anticipated trends. The occurrence of maxima for the oxides $FeO + Fe_2O_3$, $TiO_2$, $P_2O_5$, and MnO in the rocks of the transitional zone as indicated by analysis B, table 1 and figure 8, are consistent with

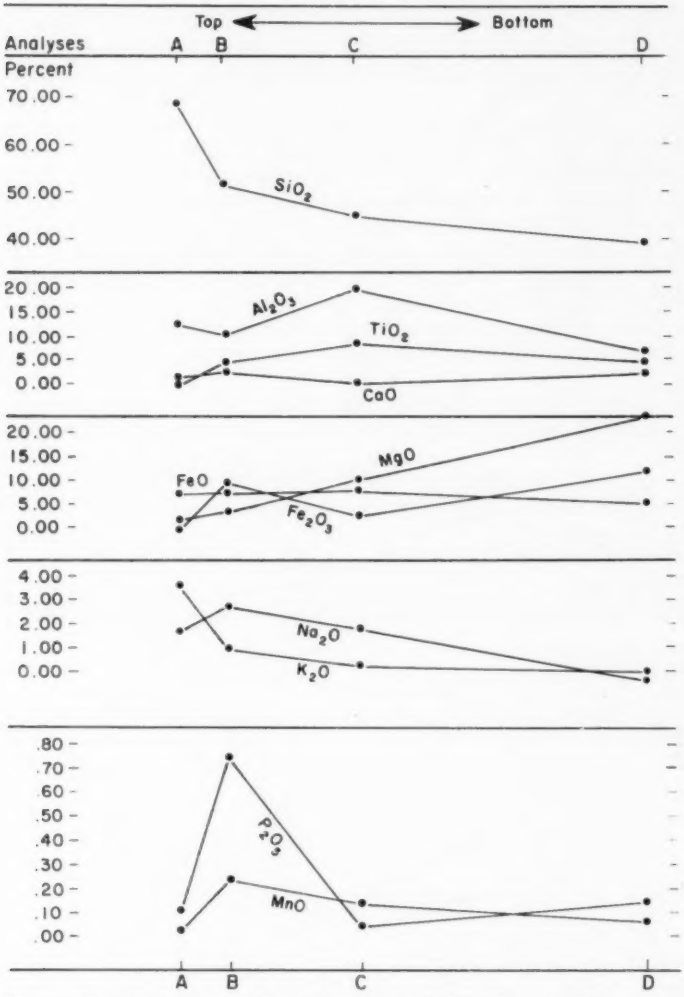



Fig. 8. Diagram showing oxide variations across the West Kiernan sill.

observations elsewhere and are considered to be chiefly caused by differentiation, at least in part by mineral fractionation (Fenner, 1929, p. 227; Collins, 1934, p. 143-148; Wager and Deer, 1939, p. 5-6; Walker and Poldervaart, 1949, p. 653; Cornwall, 1951, p. 156-158; and Hotz, 1953, p. 694).

It is important to note that only that part of the curves connecting corresponding oxides of analyses A, B, and C may represent differentiation by mineral fractionation and that the curves between analyses C and D represent a discontinuity in the series, a scattering (Bowen, 1928, p. 124), caused by the

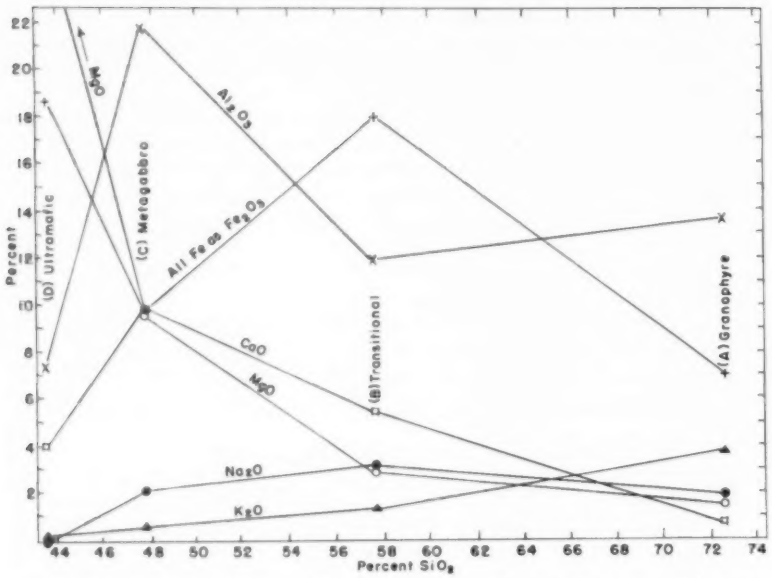



Fig. 9. Variation diagram of West Kiernan sill analyses plotted on same bases as variation diagram in figure 10. Letters in parentheses indicative analyses on table 1.

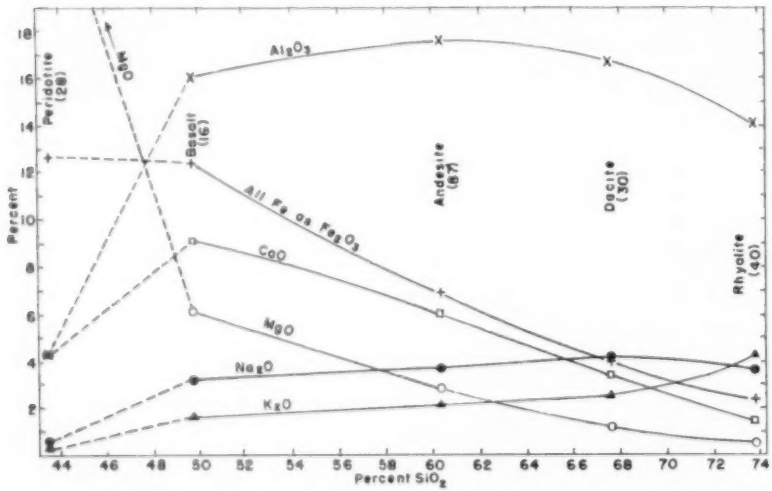



Fig. 10. Variation diagram for Daly's average basalt, andesite, dacite, and rhyolite (the full curves). In addition, dotted lines connect corresponding oxides of basalt and peridotite (Daly's average). The dotted lines show the random relation of the sorted rock (peridititte) to the curves of the average liquid line of descent (full curves). Numbers in parentheses give number of rocks averaged. (After Bowen, 1928, p. 123).

accumulation of heavy mafic minerals in the bottom of the sill. The discontinuity in the series is best illustrated when the oxides are plotted relative to silica as on figure 9. A variation diagram taken from Bowen (1928, p. 123) is given for comparison (fig. 10).

On a qualitative basis there is an obvious similarity between the two diagrams in the area of scattering—that is, between 40 and 50 percent silica. Only iron trends in an abnormal way in the Kiernan sill, but this is accounted for by the high iron content of the original magma that gave rise to an early magnetite phase which settled to the base of the sill. Wager and Deer (1939, p. 311) have suggested that if a basaltic magma contains 3.5 percent or more $Fe_2O_3$, magnetite will form as an early crystalline phase. The original Kiernan magma probably fulfilled the above condition. The fact that the ultramafic rocks contain an excess of iron and 5 percent or more of magnetite greatly enhances the probability that magnetite formed early and accumulated by gravity sorting.

Wager and Deer (1939, p. 314) have used a triangular diagram to show the differentiation trend with respect to total alkalies, FeO, and MgO, and this diagram has come into general usage (fig. 11).

The Kiernan analyses are compared on the diagram with the callucated successive liquids of the Skaergaard intrusive which Wager and Deer (1939, p. 313) regard as showing roughly the trend of strong fractionation of the average basic magma. Note that hematite is present in the norm of analysis D, table 2, and a small amount of hematite is present in the analyzed rock, and therefore point D, figure 11 probably belongs closer to point C. Point Av is

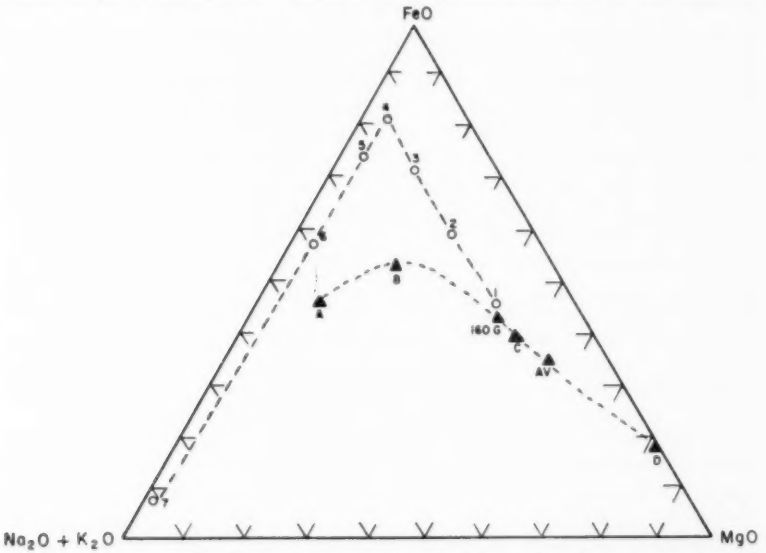



Fig. 11. Triangular diagram showing the variations in the relative proportions of the oxides FeO, MgO, and $Na_2O$ plus $K_2$). Letters A to D refer to chemical analyses on table 1; Av is the computed average for the West Kiernan sill; and 160 G is plotted from the average abalysis of 160 gabbros (Nockolds, 1954). Circles represent calculated successive liquids of the Skaergaard intrusive (Wagner and Deer, 1939, p. 314).

from the computed average composition of the sill based on the 4 new analyses and the relative areas of the various zones. Nockolds' average analysis of 160 gabbros is shown for comparison as point 160 G (Nockolds, 1954, p. 1020).

The plotted Kiernan analyses follow the classic trend—that is, upward from the MgO corner, which represents the most refractory normative components, toward the FeO corner as FeO (medium refractory components) and $Na_2O + K_2O$ (least refractory components) increase concomitantly; thence downward toward the $Na_2O + K_2O$ corner as the least refractory components increase with respect to FeO and MgO. It should be noted that the MgO – FeO part of the curve represents nearly 98 percent of the Skaergaard liquid, and over 90 percent of the Kiernan rocks. Plotting $Na_2O + K_2O$ and MgO against total iron ($Fe_2O_3 + FeO$) has the effect of raising all of the points along the curve, but does not significantly change the curve. The petrogenetic relationship characteristic of sills is clear in either case—that is, under conditions of extreme fractionation of basic magma, absolute enrichment of iron takes place until over 90 percent of the magma has crystallized, at which point the enrichment culminates, and the few percent of residual liquid crystallizes as relatively iron-lean granophyre.

## CONCLUSIONS

The early life history of the West Kiernan sill includes these phases: 1) intrusion, 2) crystallization, 3) folding, and 4) metamorphism. The last two of these were probably contemporaneous.

In phase 1) the basaltic magma forced its way between the beds of the Hemlock formation with pressure enough to raise 5,000 feet or more of superincumbent rock; Phase 2), the crystallization, is presumed to have started immediately after the intrusion. The chilled marginal rocks were first to form, and at the same time or soon after, olivine, pyroxene, and magnetite began to crystallize and to gravitate toward the bottom of the chamber. By the time the ultramafic zone was completely formed, normal gabbro had started to crystallize from the magma, and it continued to crystallize and to differentiate by mineral fractionation, until more than 90 percent of the magma had crystallized, and until the residual magma had the composition of granophyre. At approximately this stage, the rocks already formed immediately below the granophyre magma developed their distinctive mineralogy. Interstitial micropegmatite makes up an important part of these transitional rocks, and the transitional zone therefore is viewed as a zone of saturation with respect to the residual melt. The early formed minerals, pyroxene or hornblende, and calcic plagioclase, reacted with the residual melt as cooling proceeded to produce the secondary mineral suite; plagioclase-hornblende-stilpnomelane-epidote. Similar alteration may be seen in transitional rocks of unmetamorphosed sills.

The residual melt crystallized last producing the granophyric zone. As stated, crystallization of the sill probably began simultaneously at the bottom and at the top, but later crystal accumulation took place from the bottom up. This is indicated by the position of the granophyric zone, above center, and the geochemical trends in the rocks lying below the granophyric zone. The uniform texture and massive character displayed by the sill rocks, particularly those in the north part of the sill, suggest that part of the sill probably crystallized with-

out disturbance by convection currents, whereas the rhythmical layering in the south part of the sill may indicate that convection currents were active there. The layers undoubtedly represent rhythmical deposition of crystals during crystallization. According to Wager and Deer (1939) and Hess (1956, p. 450) such features are most likely related to convection currents in magma. The arguments put forward by Wager and Deer for deposition of the rhythmic layers of the Skaergaard intrusive by convection currents are most convincing, and the writer views the similar layering in the West Kiernan sill as originating in the same way, even though no feature was observed in the sill suggesting rising or descending currents of magma.

During the post-Animikie orogeny and metamorphism, the sill was folded into its present near vertical attitude and metamorphosed at temperatures commensurate with the formation of green schist facies rocks. Despite the fact that the mineralogy of the sill rocks has been almost completely changed by the metamorphism, most of the mineral alteration observed requires the addition of water only, and therefore the varying chemical composition of the rocks probably chiefly reflects chemical differences imposed by the differentiation of the magma and late deuteric alteration.

The regional metamorphism was most effective in reducing the rocks of the normal metagabbroic zone and the metadiabasic (chilled) zone to the stable assemblage actinolite-albite-epidote-chlorite. The rocks of the ultramafic and transitional zones which underwent pre-metamorphic deuteric alteration were perhaps not changed by the regional metamorphism except in degree of alteration, whereas the granophyric rocks remained nearly unaltered.

## References


Balsley, J. R., James, H. L., and Wier, K. L., 1949, Aeromagnetic survey of parts of Baraga, Iron and Houghton Counties, Michigan, with preliminary geologic interpretations: U. S. Geol. Survey, Geophysics Inv. prelim. rept.

Bastin, E. S., 1938, Hydrothermal alteration of rocks at Pigeon Point, Minnesota: Jour. Geology, v. 46, no. 8, p. 1058-1072.

Bayley, R. W., 1956, Geology of Lake Mary quadrangle, Iron County, Michigan: U. S. Geol. Survey open file rept., Mar. 8, 1956.

Blackader, R. G., 1956, Differentiation and assimilation in the Logan sills, Lake Superior District, Ontario: Am. Jour. Sci., v. 254, p. 623-645.

Bowen, N. L., 1928, The evolution of igneous rocks: Princeton, Univ. Press, 334 p.

Clements, J. M., and Smyth, H. L., 1899, The Crystal Falls iron-bearing district of Michigan: U. S. Geol. Survey Mon. 36.

Collins, W. H., 1934, Life history of the Sudbury Nickel Irruptive, Part I, Petrogenesis: Royal Soc. of Canada Trans., 3rd ser., v. 28, sec. 4, p. 123-178.

Cornwall, H. R., 1951, Differentiation in magmas of the Keweenawan series: Jour. Geology, v. 59, no. 2, p. 151-172.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: New York, McGraw-Hill Book Co., 563 p.

Fenner, C. N., 1929, The crystallization of basalts: Am. Jour. Sci., v. 18, p. 225-253.

Gair, J. E., and Wier, K. L., 1956, Geology of the Kiernan quadrangle, Iron County, Michigan: U. S. Geol. Survey Bull. 1044, 88 p.

Hess, H. H., 1956, The magnetic properties and differentiation of dolerite sills—a critical discussion: Am. Jour. Sci., v. 254, no. 7, p. 450.

Hotz, P. E., 1953, Petrology of granophyre in diabase near Dillsburg, Pennsylvania: Geol. Soc. America Bull., v. 64, p. 675-704.

James, H. L., 1955, Zones of regional metamorphism in the Precambrian of northern Michigan: Geol. Soc. America Bull., v. 66, p. 1455-1488.

Leighton, M. W., 1954, Petrogenesis of a gabbro-granophyre complex in northern Wisconsin: Geol. Soc. America Bull., v. 65, p. 401-442.

Nockolds, S. R., 1954, Average chemical compositions of some igneous rocks: Geol. Soc. America Bull., v. 65, p. 1007-1032.

Turner, F. J., and Verhoogen, Jean, 1951, Igneous and metamorphic petrology: New York, McGraw-Hill Book Co., 580 p.

Wager, L. R., and Deer, W. A., 1939, The petrology of the Skaergaard intrusion, Kangerdlugsuak east Greenland: Meddel. om Grønland, v. 105, no. 4, 335 p.

Walker, F., and Poldervaart, A., 1949, Karroo dolerites of the Union of South Africa: Geol. Soc. America Bull., v. 60, p. 151-154; 591-706.

Washington, H. S., 1917, Chemical analyses of igneous rocks published from 1884 to 1913, inclusive, with a critical discussion of the character and use of analyses (a revision and expansion of Professional Paper 14): U. S. Geol. Survey Prof. Paper 99, 1201 p.

TABLE 1

Chemical Analyses of Four Rock Samples from the West Kiernan Sill

| Specimen | (A) RB-15-53 | (B) RB-8-55 | (C) RB-32-53 | (D) RB-30-53 |
|---|---|---|---|---|
| Specific gravity | 2.68 | 2.85 | n.d. | 2.81 |
| $SiO_2$ | 69.12 | 52.2 | 44.94 | 39.01 |
| $TiO_2$ | .73 | 2.9 | .73 | 2.04 |
| $Al_2O_3$ | 12.83 | 10.8 | 19.70 | 6.56 |
| $Fe_2O_3$ | .78 | 8.8 | 1.72 | 11.49 |
| FeO | 5.85 | 7.5 | 7.40 | 5.30 |
| MgO | 1.38 | 2.7 | 8.91 | 23.84 |
| CaO | .70 | 4.8 | 9.22 | 3.57 |
| $Na_2O$ | 1.72 | 2.8 | 1.94 | 0.00 |
| $K_2O$ | 3.44 | 1.1 | .36 | .02 |
| $H_2O-$ | .09 | 3.9 | .11 | 0.66 |
| $H_2O+$ | 2.25 | | 4.56 | 7.10 |
| $CO_2$ | .70 | .37 | .45 | .08 |
| $P_2O_5$ | .11 | .76 | .09 | .19 |
| MnO | .05 | .26 | .13 | .11 |
| Totals | 99.75 | 98.9 | 100.26 | 99.97 |
| Cu, ppm | 16 | n.d. | 23 | 125 |
| Co, ppm | 0 | n.d. | 15 | 40 |
| Ni, ppm | 0 | n.d. | 72 | 390 |
| Cr, ppm | n.d. | n.d. | n.d. | 2400 |

(A) RB-15-53, sheared granophyre from 700 feet north and 750 feet east of the W1/4 corner sec. 29, T. 43 N., R. 31 W.

(B) RB-8-55, metagabbro from transition zone, SW1/4 NW1/4 sec. 29, T. 43 N., R. 31 W.

(C) RB-32-53, metagabbro, normal type, from 300 feet north and 400 feet east of the S1/4 corner sec. 29, T. 43 N., R. 31 W.

(D) RB-30-53, metaperidotite from 1350 feet north of the S1/4 corner sec. 21, T. 43 N., R. 31 W., Iron County, Michigan.

Analysts: for A, C, and D, Lucille M. Kehl, U. S. Geological Survey, 1954.
for B, Paul L. D. Elmore and Katrine E. White, U. S. Geological Survey, 1955.

TABLE 2

Norms and Modes of West Kiernan Sill Rocks

| Norms | A | B | C | D |
|---|---|---|---|---|
| Specimen | RB-15-53 | RB-8-55 | RB-32-53 | RB-30-53 |
| Quartz | 40.3 | 19.1 | 1.3 | — |
| Albite | 14.1 | 23.0 | 16.4 | — |
| Anorthite | — | 13.6 | 42.8 | 16.7 |
| Orthoclase | 20.0 | 6.5 | 2.1 | — |
| $CaSiO_3$ | — | 3.9 | — | — |
| $MgSiO_3$ | 3.4 | 6.7 | 22.2 | 47.0 |
| $FeSiO_3$ | 9.0 | 1.6 | 11.3 | — |
| $Mg_2SiO_4$ | — | — | — | 8.4 |
| Ilmenite | 1.1 | 5.5 | 1.3 | 2.0 |
| Magnetite | 1.2 | 12.8 | 2.3 | — |
| Hematite | — | — | — | 11.5 |
| Apatite | 0.2 | 1.6 | — | 0.4 |
| Corundum | 6.4 | — | 0.2 | 0.4 |
| Rutile | — | — | — | 1.0 |
| Calcite | 1.0 | — | 1.0 | — |
| | 96.7 | 94.3 | 100.9 | 87.4 |

| Modes | A | B | C | D |
|---|---|---|---|---|
| Micropegmatite | 50.0 | 10.0 | — | — |
| Microcline | 41.0 | — | — | — |
| Albite | trace | 10.0 | — | — |
| Epidote | trace | trace | 75.0* | — |
| Actinolite | — | — | 25.0# | trace |
| Tremolite | — | — | | 35.0 |
| Chlorite | 2.0 | 30.0 | | — |
| Sericite | 3.0 | — | — | — |
| Stilnomelane | trace | 46.0 | — | — |
| Biotite | trace | — | — | — |
| Serpentine | — | — | — | 60.0† |
| Talc | — | — | — | — |
| Apatite | trace | 0.5 | trace | trace |
| Sphene | — | — | — | trace |
| Leucoxene | trace | 3.5 | trace | — |
| Tourmaline | — | trace | — | — |
| Magnetite | trace | trace | trace | 5.0 |
| Calcite | 4.0 | trace | — | — |
| | 100.0 | 100.0 | 100.0 | 100.0 |

* Chiefly clinozoisite, but some albite.

\# Chiefly colorless tremolitic amphibole, but some actinolite and colorless chlorite.

† Chiefly serpentine, but some tremolite and talc.

TABLE 3

Chemical Analyses of a Transitional Rock and a Corresponding Type

| | (1) RB-8-55 | (2) W-687-4 |
|---|---|---|
| $SiO_2$ | 52.2 | 50.11 |
| $TiO_2$ | 2.9 | 3.43 |
| $Al_2O_3$ | 10.8 | 12.73 |
| $Fe_2O_3$ | 8.8 | 5.17 |
| FeO | 7.5 | 9.97 |
| MgO | 2.7 | 3.73 |
| CaO | 4.8 | 7.30 |
| $Na_2O$ | 2.8 | 2.33 |
| $K_2O$ | 1.1 | 1.76 |
| $H_2O$ | 3.9 | 2.48 |
| $CO_2$ | 0.37 | — |
| $P_2O_5$ | 0.76 | 0.47 |
| MnO | 0.26 | 0.32 |
| | 98.89 | 99.80 |

(1) RB-8-55, analysis (B) from table 1.

(2) W-687-4, analysis of transitional metagabbro in northern Wisconsin (Leighton, 1954, p. 426).

TABLE 4

Modes of Four Specimens of Granophyre from West Kiernan Sill

| | 114-52 | 115-52 | 109-52 | 109a-52 |
|---|---|---|---|---|
| Potassium feldspar (orthoclase or microcline) | 37 | 10 | | 9 |
| Micropegmatite | 57 | 80 | 62* | 75 |
| Albite | T | 7 | 33 | 8 |
| Micas (chlorite-biotite-sericite) | T | 1 | 3 | T |
| Carbonate | | 2 | .5 | |
| Ore minerals | 6 | T | 1.5 | 8 |
| Apatite | T | T | T | |

* Includes some free quartz and altered potash feldspar.

T = Trace.

[American Journal of Science, Vol. 257, June 1959, P. 431-448]

# FACTORS GOVERNING EMPLACEMENT OF VOLCANIC NECKS

A. R. McBIRNEY

Department of Geology, University of California, Berkeley, California

ABSTRACT. Previously advocated processes are found inadequate to explain the rise of volcanic necks from magmatic sources to shallow levels where they can erupt on the surface. Comparisons of heat conduction of wall rocks with heat transfers of corresponding convection systems show that circulation of fluid magma is likely to take place. Rocks at the apex and walls of the intrusion can be disrupted by thermal stresses and removed as inclusions in downward flowing currents to provide space for the volume of rising magma.

## INTRODUCTION

In an effort to arrive at a better understanding of principles involved in the formation of volcanic vents, an attempt has been made to analyse certain of their observed structural and lithological features in the light of known thermodynamic processes.

Two groups of well-exposed volcanic necks well suited for such a study are found in northwestern New Mexico and the adjacent corner of Arizona. Those of the Navajo-Hopi country have been well described by Gregory (1917), Williams (1936), Hack (1942), and Shoemaker (1956), and the necks near Mount Taylor were studied by Dutton (1885), Johnson (1907) and Hunt (1938). Geikie (1902) described another group of dissected necks in Scotland. In a less deeply eroded area of Swabia a large cluster of "diatremes" has been studied by Branco (1894) and H. Cloos (1941), and another group in southeastern Missouri has been described by Rust (1937).

The great number and variety of these igneous bodies, variable depths of exposure, and clear stratigraphic relations have made possible observations founded on abundant evidence. Despite detailed petrographic and structural studies of the necks, no satisfactory explanation has been advanced for notable features related to their emplacement.

## OBSERVED FEATURES

The necks of the Navajo-Hopi country were divided by Williams (1936) into two types. The "Hopi" type, which is also well developed in the Mount Taylor region, is characterized by buttes of massive lava, many of which show columnar jointing that is vertical near the top and diverges outward to maintain a perpendicular relation to inward dipping tuffs at the base. Underlying the tuffs, sedimentary beds are in their original flat attitude. In places where a feeding channel has been exposed it is usually composed of horizontally jointed lava and transects the sedimentary beds with little apparent displacement.

The typical "Navajo" neck, believed by Williams to be the product of deeper erosion of a more explosive volcano, is a rudely cylindrical shaft with no visibly associated lavas. Shoemaker (1956) has shown that many of these still preserve downwarped beds of tuff and sediments deposited in craters following eruption. Similarly, better preserved tuff rings or "maars" in Swabia

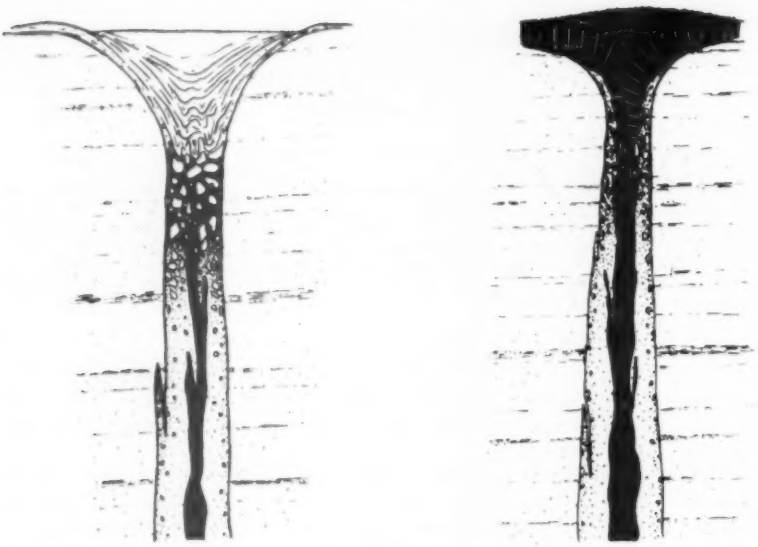


Fig. 1. Schematic vertical section through "Navajo" (left) and "Hopi" (right) type necks.

suggest that rise of the necks culminated in brief, explosive eruptions with little subsequent activity. Such features of uppermost portions of the necks appear to be related to late events in the formation of the vents, and the distinction between "Hopi" and "Navajo" types is obscure in deeply eroded necks where evidence of the manner of surface eruption has been removed.

Necks commonly vary in diameter from 10 or 20 meters to as much as 500 meters or more. Although most are generally circular in plan, those that have followed pre-existing fractures and dikes may fail to attain a high degree of symmetry. In the Mount Taylor field, lava-capped mesas show that the tops of the necks are on the order of 100 to 200 meters below the land surface upon which the lavas were erupted. In the Navajo field, erosion has removed most of the surface flows fed by the necks, but in one case in Chuska Valley a difference of elevation of about 400 meters can be measured between the base of the exposed necks and the lowest lava horizon. Plutonic and metamorphic xenoliths in Navajo necks derived from recognizable basement rocks indicate a downward extent of at least 1000 meters. In a few cases, notably at the southern end of the Mount Taylor field, rise of intrusions was arrested short of the surface.

Radial dikes are found around a few large necks but are otherwise uncommon. The shape of interior dikes is very irregular, both in plan and section. Some necks are partly surrounded by concentric shells that are more or less celery shaped and are sometimes separated from the main bodies by an intervening shell of undisturbed sediments.

Adjoining sedimentary rocks normally show no significant displacement,

PLATE 1

Chaistla, one of the larger Navajo necks.

PLATE 2

A closer view of Chaistla neck showing concentric shell structure.

PLATE 3

A typical, small Navajo neck. Dimensions about 10 x 30 meters.

distortion or compaction. Thermal effects on both the wall rocks and inclusions within the neck are generally limited to a narrow discolored zone measureable in inches.

Vertical sections through two schematic necks are shown in figure 1. In the "Hopi" type, remnants of eruptive lavas are preserved, while in the "Navajo" type the uppermost portion may be filled with inward dipping beds of tuff and interstratified, freshwater sediments. These are often underlain by a macrobreccia of sedimentary blocks dropped from wall rocks higher in the vent. In both types this upper section narrows downward and passes into a nearly straight-walled pipe, which is usually filled with more finely brecciated sediments in an igneous matrix. The core of the pipe is often made up of igneous rocks containing relatively few inclusions of shallow origin.

In plan, individual exposures of deeply eroded pipes are likely to be highly variable. What might be considered an ideal cross section of the lower portion of a neck was described by Rust (1937). The outcrop "is characterized by a ring of conglomeratic material (carrying boulders up to 3 feet across) with a diameter of about 60 feet. This outer phase is seen to encircle a central core which is about 25 feet in diameter and is free from megascopic

PLATE 4

Flow lines around a block of baked sandstone. Same neck as plate 3.

fragmental material. The contact between these two phases is gradational and some vertical shearing has occurred in this intermediate zone. The central core carries an abundance of small feldspar fragments obviously derived from disintegrated granite. Parallel alinement of long axes of feldspar fragments and of phenocrysts from the magma gives the core rock a conspicuous flow-line texture."

Sedimentary rocks of the breccia are derived mainly from adjacent sandstones and shales, but rocks of deeper origin are not uncommon. Inclusions range in size from as much as three feet or more in the outer margins to finely comminuted sand nearer the center. Rocks from deeper origins seldom exceed two or three inches in diameter and are more rounded than the large, angular blocks from upper levels. Stratigraphic relationships sometimes show downward displacements of xenoliths amounting to hundreds of feet.

In some Navajo necks, Shoemaker and Moore (1956) were able to determine the stratigraphic origin of inclusions. An excellent example is the Mule's Ear pipe (fig. 2), in which an outer column containing sedimentary fragments from beds much nearer the surface surrounds a central core with inclusions carried up from below.

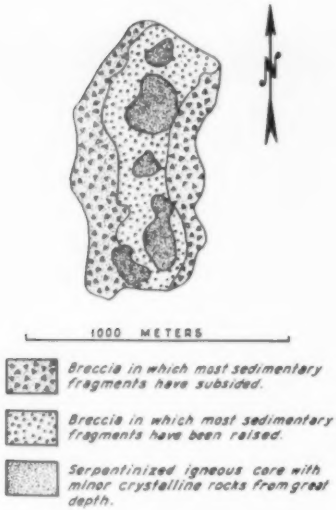

N
1000 METERS
Breccia in which most sedimentary fragments have subsided.
Breccia in which most sedimentary fragments have been raised.
Serpentinized igneous core with minor crystalline rocks from great depth.

Fig. 2. Diagrammatic plan of the Mule Ear neck in the Navajo Indian Reservation. (After Shoemaker and Moore, 1956)

Flow structure, interpreted by some as bedding, has been observed in the breccias of deeply eroded pipes of several regions. Vertical striations on wall rocks have been reported by Cloos (1941) in Swabian pipes, by Rust (1937) in Missouri, and by Shoemaker in the Navajo region.

## EMPLACEMENT PROCESSES

Among the processes advocated for emplacement of the necks three are worthy of consideration, namely, (1) explosive boring; (2) forcible intrusion with either displacement or melting of the intruded beds; and (3) churning caused by successive upward and downward surges of magma.

Explosive formation was advocated by Geikie (1902) for the vents of eastern Fife and by Branco (1894) and Cloos (1941) for the Swabian "diatremes", the breccias being interpreted as tuff that fell back into open vents following eruption. Hunt (1938) has pointed out difficulties of such an origin for breccias in the Mount Taylor necks. Fragmental sedimentary debris or pyroclastic material is practically absent from within and beneath the related lavas, which are all of massive, nonexplosive type. Explosive "boring" is hardly consistent with the restriction of breccias to the interior of conduits, some of which never reached the surface. Secondly, it is unlikely that every one of the deep, narrow pipes would have remained open to collect pyroclastic breccia at depths of a hundred meters or more in weak sediments.

Cloos (1941), after detailed study of Swabian vents, concluded that hot, tuff-laden gases of exceptionally explosive character were channeled through joint intersections to the surface and by a process of intense erosion scoured the fracture walls and detached large blocks which were brecciated by the gas

current. This process has probably been effective near the surface or in established vents of mature volcanoes but has doubtful application to depths at which joints leave little space for the volumetric increase that accompanies brecciation. Prior to eruption on the surface velocities of gas currents must be limited, and high temperatures could not be maintained at great distances from the magmatic source, especially through a disperse, reticulated fracture system. Without prior existence of a large channel, pressures at such depths could not be reduced sufficiently to produce gas velocities high enough to erode wall rocks. Confirmation of this fact is found in steam wells, such as those of New Zealand, which have been drilled to depths of over 2000 feet. Once loose fragments have been cleaned out in the initial "blow off", wells show little tendency to enlarge their diameters by erosion even when freely discharging large quantities of high temperature steam. It has been found by actual measurement (Fisher, 1955), that regardless of total depth the greatest drop in pressure and increase in velocity occurs in the uppermost 5 feet of a freely discharging well.

The flaring, upper portions of many "Navajo" type vents appear to have been formed by mild, phreatic eruptions, followed by inward slumping of the walls and extended periods of sedimentation and subsidence. Hack (1942) pointed out that the land surface at the time of eruption was "an aggrading basin of sedimentation, and was in part wet and swampy". Abuandant groundwater would tend to produce steam eruptions when intrusions reached shallow levels at which vapor pressures exceeded hydrostatic loads. Elsewhere, under drier conditions, vents were opened by upwelling of relatively non-explosive lavas carrying little sedimentary debris.

Many obstacles oppose a theory of forcible intrusion of magma under high pressure. No adequate means is offered to provide the necessary space without displacing the surrounding rocks, and the lack of distortion indicates that little such disturbance could have occurred. The slenderness of the pipes, the scarcity of radial dikes and sills, and absence of upward displacement of bedding all suggest rather weak vertical forces. Dilation of the country rock by tensional forces might produce dikes but certainly not circular or oval necks. Appeal to melting would require unusually high temperatures without any means of maintaining them. Such temperatures are not consistent with the slight degree of thermal alteration of inclusions and contacts.

The churning process described by Hunt (1938) comes closer to explaining the breccias. That pulsating forces operate in volcanic vents is an unexplained but well substantiated fact. There remain, however, the two fundamental problems of (1) disposing of breccia in order to make room for introduced igneous material and (2) supplying continued heat energy to maintain a fluid upward growth.

The only conceivable process capable of satisfying these two essential requirements would seem to be a convection system, which would perform the dual function of removing sedimentary material and bringing in hot, fresh magma. Although the emplacement process is, by its nature, one of energy disequilibrium, we may still expect it to operate within the limitations imposed by well established thermodynamic principles. If it can be shown, therefore,

that a convection system can exist under the observed conditions, the exact balance or magnitude of the disequilibrium becomes of less importance.

## ANALYSIS OF A CONVECTION SYSTEM

Since the operation of a convection system depends upon cooling which produces a difference in density between downward and upward moving columns, the problem is to show that such a difference can be created and maintained by thermal conduction through wall rocks and that this cooling can be great enough to produce the observed results.

A steady state of circulation will be considered in which a central column rises from a source of basaltic magma at temperature $T_1$, and a descending, outer column has a given average temperature $T_o$ at the outer boundary. Rates at which heat can be conducted away from boundaries of intrusions of various sizes will be compared with rates at which heat is brought up by the rising column. The rate of heat conduction must equal or exceed the rate of heat flow in the corresponding convection cell if circulation is to be maintained. Heat dissipation by escape of volatiles or any means other than conduction will be neglected. Physical values to be used are those given in table 1. Notation is given in table 2.

Ability of a country rock to conduct heat away from the boundary surface and maintain the cooling effect will be a function of its conductivity, the time elapsed since intrusion, and other factors.

Equations for the conduction of heat from long, cylindrical columns have been given by Ingersoll and Zobel (1938). The heat loss per unit height, q, is equivalent to the flux of a line source of heat, q′, at the center of the column and is expressed by:

$$q = q'e^{-r^2n^2} \quad . \tag{1}$$

Rates of heat conduction away from the cylinder by the country rock can be calculated from the equation:

$$T = \frac{q'}{2\pi rk\sqrt{\pi}} \int_{rn}^{\infty} e^{-\beta^2} d\beta \tag{2}$$

where T is the boundary temperature and $\beta$ a variable of the integral. The same equation can be used to find the temperature distribution, for a constant q, after various periods of elapsed time, as shown in diagram I.

The dashed lines of diagram II are log qH-r curves curves constructed by using equations (2) and (1) for homogeneous sandstone at an assumed boundary temperature of 450°C, heights (H) of ½, 1, and 2 kilometers, and elapsed times of ½, 1, and 2 years. It is readily seen from these curves that heat flow from a fixed boundary of given radius decreases rapidly with time, and in order to maintain a constant heat loss from the column over any period of time its radius or its height must be increased. The effect on the total heat loss of an upward growth of the column is much less than radial growth.

The problem of free convection in long, narrow cavities has been ex-

amined by Lighthill (1953). He found that convective, laminar flow is possible when the expression

$$\frac{gr^4\Theta\alpha}{\frac{\nu}{\rho}\cdot h^2\cdot H} \tag{3}$$

has a value greater than the empirical, dimensionless constant 5600.

Assuming the values in table 1 and a temperature difference ($\Theta$) of 500°, it is possible to calculate a minimum ratio $\frac{r^4}{H}$ at which free convection can exist. This is found to be about 1125. Thus, if magma rises from a depth of one to two kilometers and heat dissipation is adequate, the minimum radius

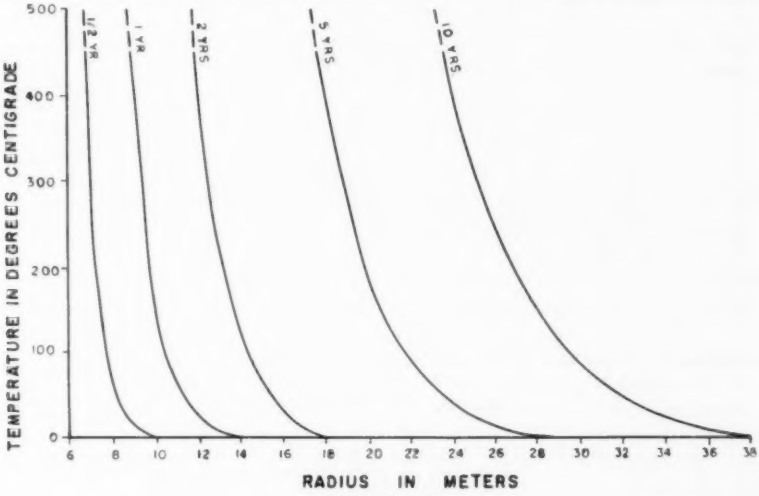



DIAGRAM 1

Distribution of wall rock temperatures at various times assuming a constant heat source.

of the intrusion is found to be of the order of 104 to 123 centimeters. Dimensions of most observed volcanic intrusions of cylindrical shape would, therefore, impose no theoretical limitation on the height to which convection systems could operate. At greater depths, where temperatures of the country rock would be higher, $\Theta$ would be correspondingly smaller, and a larger radius would be required.

Lighthill has shown that the heat transfer, Q, for a convection system can be determined from the equation:

$$Q = N_a \cdot 2\pi Hk\Theta \tag{4}$$

where $N_a$ is the Nusselt number with respect to the area of the cavity walls and can be found from:

$$N_a = .677\left[\frac{gr^4}{\frac{\nu}{\rho}\cdot h^2 H}\right]^{1/4} \tag{5}$$

Using these equations the value of Q for a basaltic magma circulating in cylindrical columns with heights of ½, 1, and 2 kilometers has been plotted against the radius of the column (solid lines, Diagram II).

From the two sets of intersecting curves of Diagram II it is possible to draw the following conclusions. Heat conducted away from the boundary by the country rock will be sufficient to cool the magma and maintain a convec-

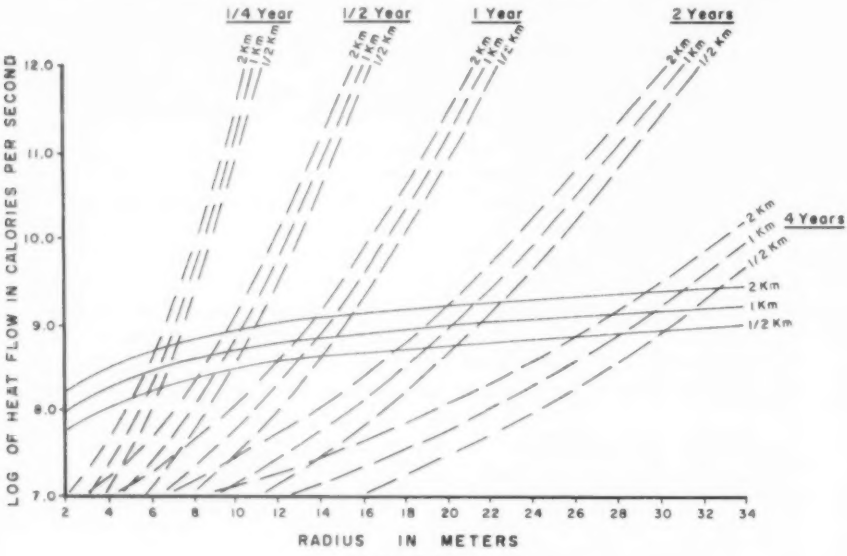



DIAGRAM II

Heat flux of convection cells of various dimensions (solid lines) and heat dissipation through walls of cylindrical cavities of the same dimensions after various periods of time (broken lines).

tion system where its value exceeds that of the convection heat transfer for a column of given height and radius. At any value of r greater than that at the intersections of the corresponding curves this condition exists.

With lapse of time conductive capacity of wall rocks diminishes as their temperature approaches that of the magma, and a larger radius is required to maintain convection. Rapid, upward growth of the column, however, would require very little outward expansion.

If the original source of magma is conceived of as a broad mass of unknown thickness, slow convection currents would be set up wherever there was an irregularity of the roof to produce a temperature differential in the fluid. Upward growth into levels of greater temperature difference would require progressively smaller radii, provided the rise were sufficiently rapid to offset the decline of heat dissipation with time.

Values assumed in the construction of these curves may vary from actual conditions by a wide margin, but the fact that the curves intersect at reasonably small values of r and that the dimensions indicated for columns are well

within the general order of magnitude of those observed in the field suggests that the physical conditions would permit, if not demand, convection currents in fluid volcanic intrusions, at least for short periods of time.

### REMOVAL OF THE INTRUDED ROCK

As an intrusion rises through the country rock it must be capable of disposing of the volume of material displaced. Since the downward-flowing current of the igneous column is in immediate contact with the walls of the growing cavity, a convenient means of removing fragmented rock is available. The country rocks at upper levels are likely, however, to be lighter than the magma, and they will move downward only if the current has a velocity greater than that with which they would tend to rise as a result of natural buoyancy.

The minimum velocity of the downward current may therefore be taken as the velocity of a rising solid in a denser fluid, which can be calculated from the equation:

$$V = \frac{2r^2(\rho i - \rho c)\,g}{v} \tag{6}$$

For example, a sandstone inclusion with a radius of 2 centimeters in a fluid having a viscosity of $2 \cdot 10^4$ poises would rise at a rate of approximately .08 centimeters per second. If the downward velocity of the current were less than this value, the size of the fragment would have to be reduced before it could be removed. Plate 4 shows an example from a Navajo neck where magma has flowed down around a block of sandstone. Smaller inclusions in the same neck appear to have been carried along with the downward current.

The magnitude of flow velocities in the breccia portion of volcanic necks is, therefore, reflected by the maximum size of fragments suspended in the necks at an appreciable distance below the roof and away from the boundary surface. A study of the size and density of xenoliths in the outer column might indicate the flow velocity in some necks were it not for the fact that the evidence of the final phase now seen is not representative of dynamic conditions during growth.

Another limiting factor affecting the capacity of the system to remove intruded rock would be the dilution effect of material having a lower density than the igneous fluid. Since the density of the outer column, $\rho_o$, must exceed that of the inner column, $\rho_i$, we can say that

$$(1-x)\,\rho_o + x\rho_c \geqslant \rho_i \tag{7}$$

where x is the percentage of country rock of density $\rho_c$ in the outer column. Since

$$\rho_o = \rho_i(1+\alpha\Theta) \tag{8}$$

the maximum permissible dilution by a sandstone country rock can be calculated and is found to range from 17 percent for a temperature difference of 500° to 7 percent for 200°. Judging from the observed breccias this range does not appear unreasonable.

Having calculated approximate values for the limiting velocity and maximum dilution we are in a position to estimate the maximum volume of country rock that the system could remove in a given time. Assuming the cross sec-

tional areas of the inner and outer columns were approximately equal and the column did not expand radially, the volume of roof material that could be removed in time t would be

$$V \cdot t \cdot x \cdot \frac{A}{2} \tag{9}$$

where V is the average downward velocity. Therefore, the maximum rate of upward growth of the column would be

$$\frac{VtxA}{2} \div At = \tfrac{1}{2}Vx. \tag{10}$$

Using a minimum downward velocity of .2 cm/sec, the maximum rate of upward advance would be about 18 meters per day for a temperature difference of 500° or 7.25 meters per day for 200°. Since other factors discussed below are likely to impose a greater limitation on upward growth than these, it seems unlikely that such a rapid rate of rise would ever be attained. It appears, therefore, that values obtained from equations (7) and (8) provide ample latitude for operation of a convection system and removal of reasonable quantities of country rock.

In the actual process of intrusion there would be considerable irregularity in almost all the variable factors that have been used in these estimates, and it is impossible to predict more than a very approximate limiting range of conditions based on assumed physical constants and average effects.

The ultimate fate of the country rock carried down by the outer column can only be guessed at, but it seems likely that a portion of it would be caught up by the rising inner column and recirculated before eventually passing out of the system or disintegrating by abrasion and thermal effects. Thus inclusions from deeper levels, such as the small, rounded xenoliths of plutonic rocks found in many necks, would be expected.

## MECHANISM OF GROWTH

The mechanism by which the igneous column rises must depend on energy derived from the magma itself, either as heat or as pressure. There can be no question of the availability of large amounts of heat. The importance of pressure, however, is less certain. The fact that lavas rise to the surface to form cones of considerable height implies pressures equalling or exceeding the weight of the adjacent rock columns. The scarcity of sills, even in well-bedded sediments under light load, suggests, however, that intrusive pressures are not great. In those rare instances when volcanic conduits have been observed reaching the surface, as at Paricutin (Gonzalez and Foshag, 1946) or at Vesuvius in 1913 (Perret, 1924), the first manifestation of the new vent has been a gradually increasing escape of gases, subsidence, and finally, after a significant interval, upwelling of lava. It appears doubtful, therefore, that in such cases volcanic intrusions grow by any important amount of "wedging" or "churning" of the roof, and one is forced to turn to heat for a potent source of energy.

Measured temperatures of lavas generally lie in the range of 1000 to 1100°C and are well below the melting points of common rock minerals. This fact is substantiated by the scant evidence of fusion in sedimentary inclusions found in the intrusions. Increase of temperature in rocks adjacent to the in-

trusion, however, will set up thermal stresses which may be responsible for breaking down the roof and walls. These stresses would probably be of two types: those due to expansion of water contained in pore spaces of the rock, and those due to expansion of the rock itself.

At shallow depths it is likely that intruded rocks will be saturated with ground water. Thermal expansion of this water in restricted pore spaces would produce within the rock large internal stresses. While these stresses cannot be calculated by any simple means, the tables of Kennedy (1950) show that pressures of well over 2000 bars would be produced by elevating the temperature of water only 200° at constant volume. The distribution of the

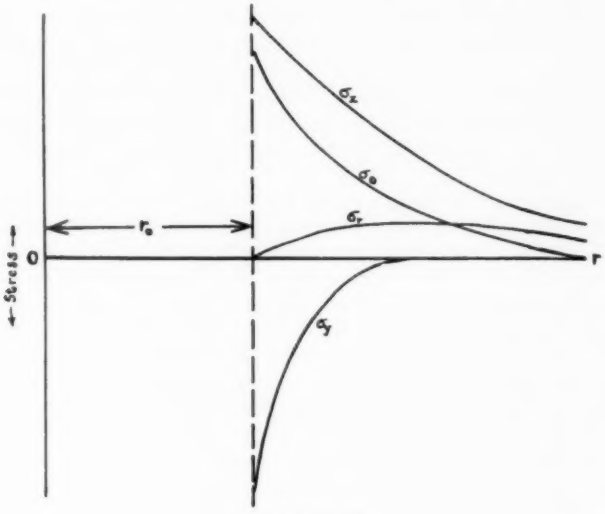



DIAGRAM IIIa

resulting stress in wall rocks would be very irregular due to jointing, bedding planes, and other variable factors, but under ideal conditions in homogeneous rock they would be a function of the porosity and temperature gradient and would have a relation to r resembling that shown as $\sigma_y$ in Diagram IIIa.

Heating of the wall rocks sets up additional stresses resulting from the effort of the rock to expand under restraining conditions. In a cylindrical surface bounding the intrusion these stresses would have three components: one, $\sigma_\theta$, tangential to the circumference; a second, $\sigma_r$, normal to the circumference; and a third, $\sigma_z$, parallel to the vertical axis. Timoshenko and Goodier (1951) have examined the problem of a long, hollow cylinder with a heat source at its center and derived equations for the thermal stress components. Adapted to the case of a cavity in an infinite body[1], their equations become:

[1] $\sigma_r$ is assumed to $= 0$ at $r_o$ and $\sigma_\theta = 0$ at $\infty$. Actually, $\sigma_r$ would equal the pressure of the magma against the wall, but the error introduced does not effectively change the distribution of the stresses.

$$\sigma_\theta = \frac{\alpha E}{1-\upsilon} \cdot \frac{1}{r^2} \int_{r_o}^{r} Trdr + \frac{\alpha E T r_o^2}{(1-\upsilon) r^2} \tag{11}$$

$$\sigma_r = \frac{-\alpha E}{1-\upsilon} \cdot \frac{1}{r^2} \int_{r_o}^{r} Trdr + \frac{\alpha E T}{1-\upsilon} \left(1 - \frac{r_o^2}{r^2}\right) \tag{12}$$

$$\sigma_z = \frac{\alpha E T (2\upsilon - 1)}{1-\upsilon} \tag{13}$$

For T, some expression involving $T_o$, r, and $r_o$ must be introduced, and under ideal conditions this would be the same as that used in equation (2). In the case of expanding radii, however, no exact expression can be applied, and it is necessary to use an arbitrary function that will reflect the steep temperature gradient next to the boundary. For such conditions, the curves for the stress components, $\sigma_\theta$, $\sigma_r$, and $\sigma_z$, will have the general shape shown in Diagram IIIa.

Since the stress due to expansion of enclosed water has three components, all equal, in the directions of $\sigma_\theta$, $\sigma_r$, and $\sigma_z$, the resultants, $\sigma_\theta + \sigma_y$, $\sigma_r + \sigma_y$, and $\sigma_z + \sigma_y$, can be found by simple algebraic addition. This has been done in Diagram IIIb.

Although it is not possible to predict absolute values for these stresses, their relative magnitudes, being dependent on physical properties of the rock and r alone, are not likely to vary widely from this relationship except in local instances.

Since the tensional strength of rocks is considerably less than the compressive strength, large negative (tensional) stresses, if they exist, are likely to produce the first failure. Examination of the curves in Diagram IIIb shows that near the boundary radial stress will have a large negative value largely dependent on the magnitude of $\sigma_y$, and tangential stress, whether positive or

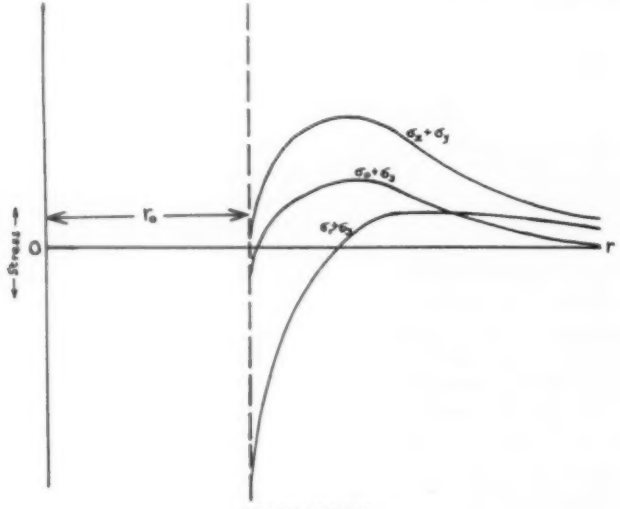



DIAGRAM IIIb

negative, will in all cases be considerably smaller in magnitude. Axial stress will have a positive value that will increase with the lithostatic load due to depth. The combined effect of these components is expressed by the strain formula:

$$\varepsilon_r = \frac{1}{E}\left[\sigma_r - v(\sigma_\theta + \sigma_z)\right] + \alpha T \qquad (14)$$

A rough calculation, based on typical values of E and $v$, shows that a relatively small value of $\sigma_r$ is required to produce negative strain, (about .5 bar at 200°, if $\sigma_\theta$ and $\sigma_z$ are neglected).

At greater depth, as porosity diminishes and the negative value of $\sigma_y$ decreases, $\sigma_\theta$ and $\sigma_z$ will begin to take on increasingly large positive values. $\sigma_z$ will increase more rapidly, due to the additional increment of static load, and it seems doubtful if $\varepsilon_r$ would ever become positive.

The effect of radial dilation would be concentric tension cracks or potential voids. These would have a negative pressure with respect to the magma, which would be likely to intrude them. Thus, a reasonable explanation if found for concentric shells observed around some of the necks.

The wall of sedimentary rock remaining between the main column and the concentric shell would be very unstable and soon would crumble. Cloos observed a type of vertical stratification in Swabian "diatremes", which he called "Kettenschichtungen" and described as chains of wall rock fragments lying close and parallel to the contact. Similar features in some Navajo-Hopi and Mount Taylor necks may represent remnants of disintegrated diaphragms between concentric shells and the main intrusive bodies.

The area of most intense heating of intruded rocks must be focused at the apex of the rising column, where hot gases are released and exothermal reactions may occur. If the temperature of sandstones or other quartz-rich rocks is raised above 573° a phase inversion of the quartz produces stresses, which would be added to thermal stresses tending to cause spalling of roof rocks. A practical application of this process is the boring of blast holes in taconite deposits by "jet piercing" (Antonides, 1958). By concentrating a jet of burning acetylene and steam against the bottom of the hole the rock is shattered and fragments are carried to the surface by the rising gas stream.

Once detached, blocks would continue to break down. Heated surfaces would be scaled off by thermal stresses in the outer layer, much in the manner of spheroidal weathering of jointed igneous rocks, thus producing the rounded forms and fresh surfaces of plutonic and metamorphic xeonliths in many necks.

## CONCLUSIONS

The combined effect of thermal stresses in roof and wall rocks tends to detach blocks of irregular size and shape. If the blocks have a lower density than the igneous medium, they will remain in the upper levels of the neck until they are small enough to be carried down by circulating currents.

Heat losses at the boundary and escape of gases produce a density differential to maintain convection currents as long as heat dissipation is greater than the minimum heat flux of the convection system. Rates of heat conduc-

tion away from contacts can be maintained as long as the igneous body is growing, either vertically or radially.

Growth will tend to produce cylindrical shapes, since a circular boundary is the most efficient natural conduction surface. Irregular shapes induced by structural features of the country rock tend to be reduced by higher thermal stresses in inward projecting surfaces.

Growth of an intrusion may be arrested short of the surface by any of a number of factors. If growth is too slow to maintain conductive heat losses the rate of convection will decline. On the other hand, too rapid growth may cause the downward-flowing column to be overloaded with fragments of low density, and the convective force, which is dependent upon a density differential, will be nullified.

Quantitative estimates, though very inexact, suggest that emplacement of volcanic necks is a process which, if it is to function at all, must proceed at a rapid rate.

Conditions governing the deeper origins of volcanism and mechanisms by which magmas are mobilized constitute problems distinct from these shallower phenomena. Similarly, factors controlling activity of volcanoes after their initial eruption on the surface must be considered apart from the intrusion of necks, since they involve many radically different conditions.

ACKNOWLEDGMENTS

Without the valuable suggestions and penerating criticisms of Dr. Jean Verhoogen this study could never have been completed. His assistance in finding methods of attacking various mathematical aspects of the problem, as well as the many hours of stimulating discussion, are deeply appreciated. Among the many persons who offered helpful criticism special thanks is due to E. M. Shoemaker of the Geological Survey. Even when not in agreement with ideas I have proposed, his dissenting views, supported by an intimate knowledge of the Navajo-Hopi necks, served to reveal several flaws in the reasoning or inadequacies of the presentation. For reviewing the manuscript and suggesting important improvements I am indebted to my friend R. E. Wilcox of the Geological Survey's volcanological section. Finally, I am pleased to be able to express once more my gratitude to Dr. Howel Williams, not only for his criticism of the manuscript but equally for his friendly counsel and encouragement.

TABLE 1

Physical Constants

| | Basaltic magma | Sandstone |
|---|---|---|
| $\alpha$, Coefficient of thermal expansion | $0.3 \cdot 10^{-4}$ (cubic) | $0.1 \cdot 10^{-4}$ (linear) |
| c, Specific heat, cal·gm$^{-1}$·deg$^{-1}$ | 0.335 | 0.200 |
| $h^2$, Diffusivity, cm$^2$·sec$^{-1}$, $(k/c\rho)$ | $5.3 \cdot 10^{-3}$ | $8.7 \cdot 10^{-3}$ |
| k, Thermal conductivity, cal·cm$^{-1}$.sec$^{-1}$, deg$^{-1}$ | $4.8 \cdot 10^{-3}$ | $4.0 \cdot 10^{-3}$ |
| $\rho$, Initial density, gm·cm$^{-3}$ | 2.7 | 2.5 |
| E, Modulus of elasticity, dyne·cm$^{-2}$ | — | $2.55 \cdot 10^{11}$ |
| $v$ Poison's ratio | — | 0.25 |
| v, Viscosity, poises | 15,000 | — |

TABLE 2

Notation

$\alpha$ — Coefficient of thermal expansion
A — Cross sectional area, $cm^2$
c — Specific heat, $cal \cdot gm^{-1} \cdot deg^{-1}$
E — Modulus of elasticity, $dynes \cdot cm^2$
$\varepsilon_r$ — Radial strain
g — Acceleration of gravity, 980 $cm \cdot sec^{-2}$
H — Vertical height, cm
$h^2$ — Diffusivity, $cm^2 \cdot sec^{-1}$
k — Thermal conductivity, $cal \cdot sec^{-1} \cdot cm^{-1} \cdot deg$
$N_u$ — Nusselt number with respect to the cavity walls
Q — Heat transfer of convection cell, $cal \cdot sec^{-1}$
q — Heat flow across boundary, $cal \cdot sec^{-1} \cdot cm^{-2}$
q′ — Heat flux of equivalent line source, per unit H, $cal \cdot sec^{-1} \cdot cm^{-1}$
n — $\frac{1}{2}(h^2t)^{-\frac{1}{2}}$, $cm^{-1}$
$\nu$ — Poisson's ratio
r — Radius, cm
$r_i$ — Radius of inner column, cm
$r_o$ — Radius of boundary, cm
$\sigma_r$ — Radial thermal stress, $dynes \cdot cm^{-2}$
$\sigma_\theta$ — Tangential thermal stress, $dynes \cdot cm^{-2}$
$\sigma_z$ — Axial thermal stress, $dynes \cdot cm^{-2}$
$\rho$ — Density, $gm \cdot cm^{-3}$
T — Temperature, degrees Centigrade
$T_i$ — Temperature of magma at its source, degrees Centigrade
$T_o$ — Temperature of boundary, degrees Centigrade
t — Elapsed time since intrusion, sec
$\Theta$ — Temperature difference between rising and descending columns, °C.
V — Velocity, $cm \cdot sec$
$\nu$ — Viscosity, poises
x — Proportion of country rock in magma

REFERENCES

Antonides, L. E., 1958, Jet piercing—the miner's rocket: Eng. Mining Jour., v. 159, p. 103-107.
Branco, W., 1894, Schwabens 125 Vulkan-Embryonen: Stuttgart.
Cloos, H., 1941, Bau und Tatigkeit von Tuffschloten: Geol. Rundschau, v. 32, p. 709-800.
Dutton, C. E., 1885, Mount Taylor and the Zuni Plateau: U.S. Geol. Survey Sixth Ann. Rept., p. 165-182.
Fisher, W. M., 1955, Production of steam from drill holes at Wairakei: New Zealand Dept. Sci. and Ind. Research, Bull. 117, p. 75-102.
Geikie, A., 1902, The geology of Eastern Fife, Geol. Survey, Scotland, Mem., p. 200-283.
Gonzalez, J. & Wm. F. Foshag, 1946, The birth of Paricutin: Smithsonian Inst. Washington, Pub. no. 3878, p. 223-234.
Gregory, H. E., 1917, The geology of the Navajo country: U.S. Geol. Survey Prof. Paper 93.
Hack, J. T., 1942, Sedimentation and volcanism in the Hopi Buttes, Airzona: Geol. Soc. America, Bull. v. 53, p. 335-372.
Handbook of Physical Constants, 1942: Geol. Soc. America Spec. Paper 36.
Hunt, C. B., 1937, Igneous geology and structure of the Mount Taylor volcanic field: U.S. Geol. Survey Prof. Paper 189-B.
Johnson, D. W., 1907, Volcanic necks of the Mount Taylor region, New Mexico: Geol. Soc. America, Bull., v. 18, p. 303-324.
Kennedy, G. C., 1950, Pressure-volume-temperature relations in water at elevated temperatures and pressures: AM. JOUR. SCI., v. 248, p. 540-564.

Ingersoll, L. R. and O. J. Zobel, 1948, Mathematical theory of heat conduction: Boston, Ginn and Co.
Lighthill, M. J., 1953, Theoretical considerations on free convection in tubes: Quart. Jour. App. Math., v. 6, pt. 4, p. 398-439.
Perret, F. A., 1924, The Vesuvius eruption of 1906: Carnegie Inst. Washington.
Rust, G. W., 1937, Preliminary notes on explosive volcanism in southeastern Missouri: Jour. Geology, v. 45, p. 48-75.
Shoemaker E. M. and H. J. Moore, II, 1956, Diatremes on the Navajo and Hopi reservations: U.S. Geol. Survey Trace Elements Inv. Rept. 640, p. 197-203.
Shoemaker, E M., 1956, Occurrence of uranium in diatremes on the Navajo and Hopi reservations, Arizona, New Mexico, and Utah: U.S. Geol. Survey Prof. Paper 300, p. 179-185.
Timoshenko, S. and J. N. Goodier, 1951, Theory of elasticity: New York, McGraw-Hill.
Williams, H., 1936, Pliocene volcanoes of the Navajo-Hopi country: Geol. Soc. America Bull., v. 47, p. 111-171.

[American Journal of Science, Vol. 257, June 1959, P. 449-457]

# SHELL DUNES OF THE SONORAN SHORE

RONALD L. IVES

2075 Harvard Street, Palo Alto, California

ABSTRACT. The Great Sand Dunes along the northeast shore of the Gulf of California extend from north of Yuma, Arizona to south of Puerto Peñasco, Sonora, but their composition changes abruptly just west of the Pinacate Peaks, of Sonora. To the north, the dunes are composed largely of rock fragments and are in part reworked Colorado River sediments; to the south, they are composed largely of shell fragments, derived from recently elevated bottom marls, geologically, and probably archaeologically, recent.

## INTRODUCTION

The Great Sand Dunes separate the Colorado River from the Imperial Valley to the north and then extend southeastward for almost 100 miles along the Sonoran shore of the Gulf of California, to just south of the seaport of Puerto Peñasco, Sonora, Mexico. (figs. 1 and 2). They were discovered by

Fig. 1. Index map showing general location of area.

early Spanish explorers, probably by both Hernando de Alarcón and Melchior Díaz in 1540-41. They are not mentioned, however, in the Munchausen-like reports of Fray Francisco de Escobar, chronicler of the Oñate expedition to Yuma in 1603.

During the last decade of the 17th and the first decade of the 18th century, the Sonoran portion of the dunes was thoroughly explored and mapped by E. F. Kino and companions (Kino, ms), and they are plainly indicated, in their correct location, as "Medanos de arena" on Kino's famous map "Passo por Tierra a la California", dated 1701.

The California portion of these dunes, often known as the Yuma Desert, was mapped and described by members of the Juan Bautista de Anza party (1774-1776), and the trail from the Yuma Crossing to Mission San Gabriel further described by Don Pedro Fages (1781).

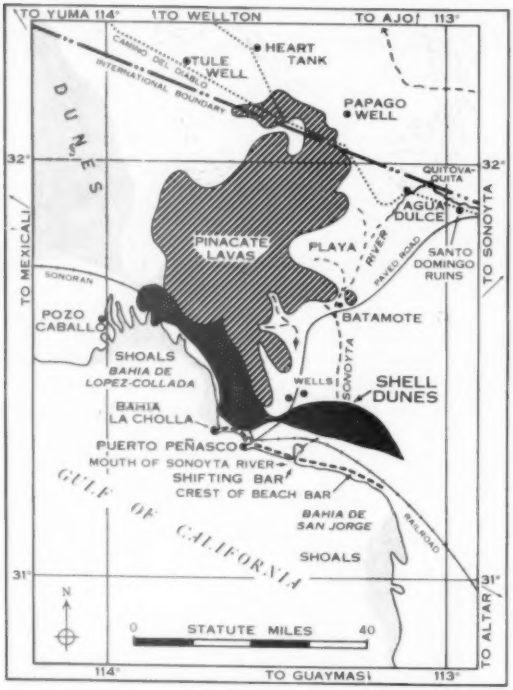



Fig. 2. Map of the Sonoran shore area southwest of Sonoyta, Sonora, showing location of shell dunes and other salient features.

Early in this century the Yuma dunes were described by Mendenhall (1909a, 1909b) and the Sonoran dunes were studied in detail by Karl Lumholtz (1912) and Alberto Celaya. The writer obtained additional information on the Sonoran dune area from the works of Godfrey Sykes (1937, Hornaday, 1908), from his son, Glenton Sykes, of Tucson, Arizona (personal communications, 1946-58), and from Alberto Celaya, of Sonoyta, Sonora (personal communications, 1931-1958).

## FIELD INVESTIGATIONS

Prior to about 1935, the whole dune belt, from north of Yuma to east of Puerto Peñasco, was thought to be formed of the same materials—the "sweepings" of the Colorado desert and drainage—and to have the same age and genesis. Field work in and around the Pinacate lava flows showed, however, that the dunes there differ notably from those farther northwest in appearance and color and even in the sound of footsteps upon them and the walking pace across them. Alberto Celaya, of Sonoyta, and several other experienced desert travellers, reported further that the various beach and pozo springs along the Gulf shore differ widely in color, taste, and toxicity, although all apparently have the same general water source. Field analyses showed that the northwest-

ern dunes have a very small calcium carbonate content, usually less than 10 percent, except in the vicinity of ancient shorelines (Sykes, 1937, p. 121, figs. 30, 33, 34; Blake, 1914; Ives, 1951), whereas the southeastern dunes are com-

PLATE 1

A. Dunes seen from the summit of Punta Peñasco. Peak on left horizon is Pinacate Peak, a quiescent volcano. Peak on right horizon is Sierra Blanca, composed of white granite. Perspective here is somewhat distorted by mirages.

B. Eastern margin of dunes, west of Sierra Blanca, where the dunes form a veneer over an outlier of the Pinacate lava flows.

posed largely of calcium carbonate, with local admixtures of volcanic ash and gypsum from obvious local sources. The volume of calcareous material in these dunes, after making a generous allowance for buried lava flows and older terrain features, is considerably more than one cubic mile.

Sand at the western edge of the calcareous dunes, when moistened by the very infrequent rains of the region, acquires an all-pervading "stockyards" odor, which continues until it is again dry. The same odor can be produced by moistening the dune sand with fresh, but not salt, water. Several "outsiders" who have travelled in this area report that "everything infects" here. Local residents are apparently immune to this difficulty.

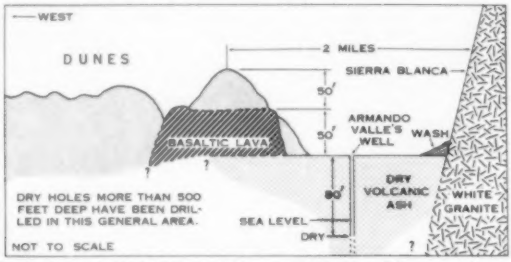



Fig. 3. Conditions in sand-free corridor west of Sierra Blanca and east of dunes.

## COMPOSITION OF DUNES

The general composition of the dune sand was determined from samples collected by the writer in the field, augmented by material collected for the writer by Alberto Celaya, Jr., of Puerto Peñasco, Sonora.

At the northwestern tip of the calcareous dunes, roughly west of Pinacate Peak, the sand contains 70 to 75 percent calcium carbonate, 25 to 20 percent rock fragments (quartz, mica, feldspar, and "black sands"), and the rest largely organic matter, in which pollen, seeds, plant fibers, rodent dung, and muscle fibers (?) can be identified. The calcium carbonate is almost entirely unfossilized shell fragments ranging in size from about 2 mm to less than .01 mm. No very fine material was found in any sample, however.

Sand collected on the east side of Bahia La Cholla contains about 85 percent shell fragments, 5 percent rock fragments, and 10 percent organic material, which here includes a few fish scales.

East of Punta Peñasco, the basalt upland sheltering the community of Puerto Peñasco, the dune sand contains varying amounts of basalt fragments, of local origin, and up to 20 percent gypsum, which was traced to thin gypsiferous lenses between the lava flows and under the beach boulders atop the 300-foot upland. Eastward, the amount of basalt and gypsum declines rapidly, becoming too small to measure about two miles to the east.

Most of the larger components of the calcareous dunes are certainly clam shell fragments, probably of the species *Chione cancellata*; at least they do not differ significantly from undoubted *Chione cancellata* fragments from the nearby shore. If these dunes are composed largely of *Chione cancellata* fragments, as seems probable, they are the remains of not less than $10^{15}$ full-grown individuals!

## DUNE CONSTRUCTION

These shell dunes lack the extreme *barchane* structure present from Yuma to Pozo Caballo, where the dominant wind is from the west. The records of the Servicio Meteorologico Mexicano, and direct field observation (Ives, 1949), show that the summer daytime wind in this part of the Sonoran Desert is westerly, being a sea breeze augmented by a gradient wind blowing tangentially into a local thermal low, which occupies the Pinacate lava flows for most of the year. This wind daily moves great quantities of sand inland from the beaches. The nocturnal land breeze of the summer season is usually present, but is ordinarily too weak to move sand.

During the hotter months of the year, roughly from April through October, dune material is periodically rearranged by dust devils, which carry loose material skyward at rates of up to a ton a minute and return the coarser particles to the general area of their origin. Finer "air float" material is carried upward to heights of at least 3,000 feet, and drifts eastward from the dune area into the Sonoyta River Valley, where it is deposited as a sort of calcareous loess. In those parts of the dunes which have poor subdrainage (as where there is a lava bed at no great depth) dust devil action is inhibited by scattered clumps of desert grasses, such as *Hilaria rigida* (pl. 1B). The general course of dust devils in this area is from southeast to northwest (right to left in pl. 1A), and from twelve to fifteen can be observed on an average summer day.

In winter, the land-sea breeze cycle is very weak, but the winter "norte" winds (Ives, 1949), coming from the northwest, cause considerable changes in dune configuration. These winds are cold, damp, and violent, and carry much dune material southeastward toward and into the Sonoyta Valley. Southeasterly dune growth there has kinked the course of the river eastward by several miles in geologically recent time, and perhaps since 1698.

## SOURCE OF DUNE MATERIAL

Because these dunes are growing rapidly, it is easy to determine that their source is the beaches and bars along the shore of the Gulf of California. These are composed largely of shells and shell fragments, as are also a number of ancient raised beaches (Ives, 1951). Many of the beaches are growing, particularly the Playa Hermosa, between Puerto Peñasco and Bahia La Cholla (pl. 2A).

At the west end of the Playa Hermosa, near Bahia La Cholla, a large expanse of rock-hard calcareous material is exposed at all tides. This is composed of whole shells and fragments of shells, all of forms still living in the Gulf, and contains considerable amounts of organic matter, some of it still actively decaying. This material is identical to the hard bottom found in many parts of the Gulf of California, usually below low tide level. It is locally called *coral*, *coquina*, *piedra de coral*, and *piedra del mar*. This material is substantially identical to the *bottom marl* and *seastone* that is quarried by divers in the waters peripheral to the Arabian Peninsula. All the local names applied to this rock are reasonable except *coral*, which is not present in it. The rock most nearly resembles the coquinite of the Florida Keys area, except for the lack of coral content.

This calcareous deposit is almost certainly bottom material which has been

PLATE 2

Beach Structures

A. Playa Hermosa, about six miles west of Puerto Peñasco, at half tide. Elevation on horizon is Punta Peñasco, composed largely of basalt: community at its foot is Puerto Peñasco.

B. Recently-elevated bottom materials, a mixture of shell fragments and organic matter, at west end of Playa Hermosa. Wind action strips away finer fragments of this material, keeping the surface free of sand.

uplifted in geologically recent time. Local evidence of this uplift is clear. The "25-foot" shoreline, present along most of the Sonoran shore of the Gulf of California, here stands about 60 feet above MSL, where at Puerto Peñasco and on the east side of Bahia La Cholla it has its nominal elevation. As this shoreline is attributed to the climatic optimum (Ives, 1951), about 6,000 years ago, the uplift is probably younger, a conclusion compatible with the modern fauna present in the uplifted bottom material and the large amount of included partially-decayed organic matter.

The possibility that this uplift took place during the last century or so, changing Pelican Point from an island to a tombolo, was suggested by a number of local legends and one map. Checking the legends disclosed that the local term "in the time of my grandfathers" is temporally indefinite, referring not to a period perhaps 75 years ago, but to any time from about 50 years ago back into the "time of the old ones" (Hohokam), which ended perhaps 1200 years ago (older times are designated "cuando era muchaco el buen Dios"). Lumholtz' map (1912) shows Pelican Point as an island, but as a sequence of other maps, dating from about 1701 (Kino) to the present (H. O. Chart 619), and including the 1908 map by Sykes (Hornaday, 1908), show the peninsula with reasonable correctness, it appears that the Lumholtz map is in error,[1] and that the region has been a peninsula for at least two and a half centuries.

Remnants of bottom marl are widely distributed over the Sonoran shore between present sea level and the old 25-foot shoreline, and the same material continues below low tide level, as shown by a large number of soundings and by several descents to depths down to 50 feet. These bottom deposits form an impervious seal, which effectively prevents sea water from entering the dry sands of the mainland, so that wells drilled along the beach in the vicinity of Puerto Peñasco are literally "dusters" to depths exceeding 500 feet.

## SUMMARY AND CONCLUSIONS

The shell dunes of the Sonoran shore of the Gulf of California are a discrete geological entity, having the same dynamic causes as the quartzose dunes farther to the north but a different source of material supply and hence a different chemical composition. Dunes west of Yuma, Arizona, are derived by aeolian erosion of local wash plus Colorado River sediments in the Imperial Valley region. Those along the extreme northern shore of the Gulf of California, roughly from Adair Bay to near Yuma, are derived directly from Colorado River sediments. The shell dunes, in contrast, consist largely of fragments of shells of *Chione cancellata*, derived from ancient and modern shell deposits.

These shell dunes are geologically recent, as they overlie in part the most recent extensive lava flows of the Pinacate series, which are not older than Pleistocene, and may be of post-Wisconsin age. The present format of the dunes dates from the attainment of the present sea level, which took place consider-

[1] Checking of diaries discloses that Lumholtz made his observations from Cholla Water, roughly 8½ miles northeast of Pelican Point. At that distance, under statistically normal conditions of curvature and refraction, the bars joining Pelican Point to the mainland are only marginally visible. Under refractive conditions prevalent here in winter, with the line of sight passing through cold air over warm water, the bars are considerably below the visible horizon of 1910, he could not see these bars, and hence did not report them.

ably less than 6,000 years ago, but surely prior to 1540.

Major present sources of dune material are the shell beach at Playa Hermosa and the upraised bottom material at its west end. As the volume of material available for wind erosion here is large, the shell dunes will remain active for some centuries, barring major changes in levels or modification of regional wind pattern.

ACKNOWLEDGMENTS

During the course of field work in the Sonoran Desert, which has extended from 1928 to the present, the writer has received useful information and valuable field assistance from a large number of people, most of them residents, or former residents, of the borderland desert area; and especially from Alberto Celaya, Sr., of Sonoyta, Sonora; Alberto Celaya, Jr., of Puerto Peñasco, Sonora; Ygnacio Quiroz, of Bahia La Cholla, Sonora; Armando Valle, of Puerto Peñasco, Sonora; Caesario Mota, of Sonoyta, Sonora; the late Jeff Milton of Tombstone, Arizona; the late José Juan Orozco, of Quitovaquita, Arizona—Sonora; the late D. T. MacDougal, of Carmel, Calif.; the late Godfrey Sykes, of Tucson, Arizona; and his son, Glenton Sykes, of the same city. Analyses of samples, performed by the Instituto de Geologia de Mexico, through the courtesy of Drs. A. R. V. Arellano, Rodolfo del Corral and Alberto Obregón Pérez, are here gratefully acknowledged. Field work was expedited, on many occasions, by the excellent cooperation given by the Mexican civil and military authorities.

REFERENCES

Alarcón, Hernando de, ms. ca. 1541, De lo que hizo por la mar Hernando de Alarcón, ——. Printed version *in* Herrera, Antonio de 1601, Historia general de los hechos de los castellanos en las islas y tierra firme del mar oceano, Dec. VI, lib. IX, cap. XIII, Madrid. English version *in* Hammond, G. P., and Rey, Agapito, 1940, Narratives of the Coronado expedition: Albuquerque, Univ. New Mexico Press, p. 124-155.

Anza, Juan Bautista de, mss. ca. 1774-76; see Bolton, H. E., 1930, Anza's California expeditions, 5 vols. internally indexed: Berkeley, Univ. Calif. Press.

Blake, W. P., 1914, The Cahuilla basin and desert of the Colorado: Carnegie Inst. Wash., Pub. 193, p. 1-12.

Diaz, Melchior, 1540-41; see Winship, G. P., 1922, The journey of Coronado: New York, Allerton Book Co., p. 26-30, 58-59.

Escobar, Francisco de, ms. ca. 1603, Relación que hizo Fray Francisco de Escobar ——, Archivo General de Indias, Seville, Audiencia de Mexico, legajo 20. English translation *in* Hammond, G. P. and Rey, Agapito, 1953, Don Juan de Oñate, colonizer of New Mexico: New Mexico Press, Albuquerque, Univ., p. 1012-1031.

Fages, Pedro, 1781, Diario de Pedro Fages, ms. in Robert E. Cowan collection, Bancroft Library, Berkeley, Calif. English translation *in* Priestley, H. I., 1913, The Colorado River campaign, 1781-1782, Diary of Pedro Fages: Acad. Pacific Coast Hist. Pubs., v. 3, no. 2, p. 135-223.

H. O. Chart 619, see Hydrographic Office Pub. 84, 1951, Sailing directions for the west coasts of Mexico and Central America, Ninth Ed. sects. 5-1 to 5-9: Washington, Government Printing Office.

Hornaday, W. T., 1908, Campfires on desert and lava: New York, Chas. Scribner's Sons, map on p. 22.

Ives, R. L., 1949, Climate of the Sonoran desert region: Assoc. Am. Geographers Ann., v. 39, p. 134-187.

——————, 1951, High sea levels of the Sonoran shore: Am. Jour. Sci., v. 249, p. 216-223.

Kino, Eusebio Francisco, mss. ca. 1701, These manuscripts, consisting of letters, trip reports, and maps, and contemporary official copies and translations thereof, number considerably more than 1,000, and are distributed among more than a dozen libraries and archival files. Dependable English translations of a majority of these, with adequate annotation and indexing by date, place, and context, and including supporting

documents by Kino's field companions and ecclesiastical superiors, are contained in Bolton, H. E., 1919, Kino's historical memoir of Pimeria Alta, 2 vols.: Cleveland, Arthur H. Clark Co.; Bolton, H. E., 1936, Rim of Christendom: New York, Macmillan; Burrus, E. J., 1954, Kino reports to headquarters: Rome, Institutum Historicum Societatis Jesu; Dunne, Peter M., 1952, Black robes in Lower California: Berkeley, Univ. Calif. Press; Karns, H. J., 1954, Unknown Arizona and Sonora (Manje's "Luz de Pierra Incógnita"): Tucson, Arizona, Silhouettes.

————, map, 1701, Passo por tierra a la California, y sus confinantes nuevas naciones y nuevas misiones de la compania de Jesus en la America septentrional, 1701. Best version published 1707 in Nova litteraria Germaniae aliorunque Europae regnorum annu MDCCVII. Collecta Hamburgi. Lipsiae et Francofurte apud Christianum Liebezeit. Modern printed version in Bolton, H. E., 1936, Rim of Christendom, op. cit. p. 160. Clear tracing in Ives, R. L., 1953, Adam Aigenler's field manual: Jour. Geog., v. 52, no. 7, p. 291-297. Error analysis in Ives, R. L., 1953, "California no es ysala," Am. Cath. Hist. Soc. Records, v. 64, p. 189-198.

Lumholtz, Carl, 1912, New trails in Mexico: New York, Scribner, p. 276-280, map in pocket.

Mendenhall, W. C., 1909a, Ground waters of the Indio region, Cal. with a sketch of the Colorado desert: U. S. Geol. Survey Water Supply Paper 225.

————, 1909b, The Colorado desert: Natl. Geog. Mag., v. 20, p. 681-701.

Sykes, Godfrey, 1937, The Colorado Delta: Carnegie Inst. Wash. Pub. 460.

[American Journal of Science, Vol. 257, June 1959, P. 458-460]

# EXAMPLES OF DEPARTURE FROM THE GAUSSIAN IN GEOMORPHIC ANALYSIS

WILLIAM F. TANNER

Geology Dept., Florida State University, Tallahassee, Florida

ABSTRACT. Many cumulative frequency plots of geomorphic data are Type I or Type IV curves in the Pearson system. These depart markedly from the Gaussian distribution. The departures can be investigated quickly by plotting the median elevation (in percent) against a percentile or decile skewness measure. A more precise plot can be obtained from skewness and kurtosis measures based on moments.

A representative geomorphic profile of a given area may be obtained by plotting cumulated elevations on ordinary probability paper. This may be done for either grid intersections or elevation peaks. The peaks used are the highest points falling within squares of a grid, one maximum elevation from each subdivision. Where both intersections and peaks are plotted (two distinct curves, on one sheet of probability paper), a clear picture of the gross landform emerges (see, for example, Tanner, 1954). The plot of intersections-only, although not quite so useful, is nevertheless a valuable tool.

The Gaussian (or "normal") distribution plots on probability paper as a straight line. The geomorphic profile is commonly not Gaussian, however, and the departures are, in many instances, quite marked. Analysis of a number of curves of this type reveals that many of them fit Pearson's Type I or Type IV (numbering according to Kendall, 1947). Three examples of the Type IV curve are presented herewith (fig. 1). Since these curves depart so sharply from the Gaussian, it becomes of interest to determine the extent of departure.

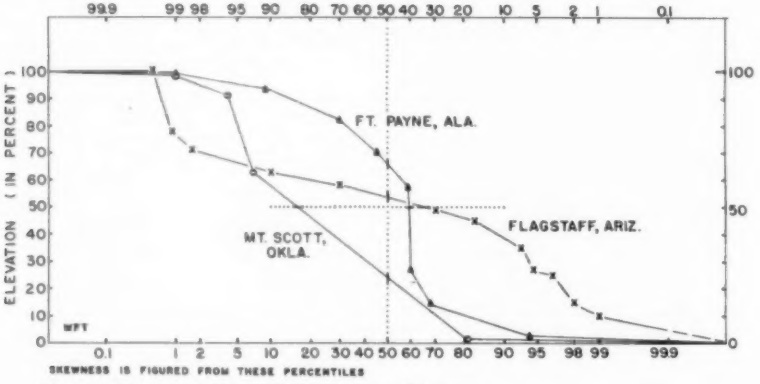

99.9 99 98 95 90 80 70 60 50 40 30 20 10 5 2 1 0.1
ELEVATION (IN PERCENT)
100 90 80 70 60 50 40 30 20 10 0
FT. PAYNE, ALA.
FLAGSTAFF, ARIZ.
MT. SCOTT, OKLA.
WFT
0.1 1 2 5 10 20 30 40 50 60 70 80 90 95 98 99 99.9
SKEWNESS IS FIGURED FROM THESE PERCENTILES

Fig. 1. Three examples of Pearson's Type IV curve.

Fortunately an evaluation of departure from the Gaussian can be based on those statistical parameters generally calculated from the first four moments about the origin. These are (Hoel, 1947):

First: the mean, a measure of central tendency.

Second: the standard deviation, a measure of variation or dispersion.

Third: skewness, a measure of asymmetry. This varies from a large positive number for regions in early, simple youth (see fig. 2) to a large negative number for regions in simple old age (see fig. 2). The Gaussian distribution has zero skewness, and the non-Gaussian generally some other value.

Fourth: kurtosis, a measure of peakedness. The Gaussian distribution has the kurtosis, $K = 3$; the non-Gaussian generally varies from this. High numerical values suggest a curve like that in the top center of Fig. 2; low values, a curve like that in the bottom center of Fig. 2.

Skewness can be plotted against kurtosis (fig. 2) to indicate how one region compares with others. In the examples, elevation has been expressed in percent of maximum relief, in order to simplify calculations (i.e., for a

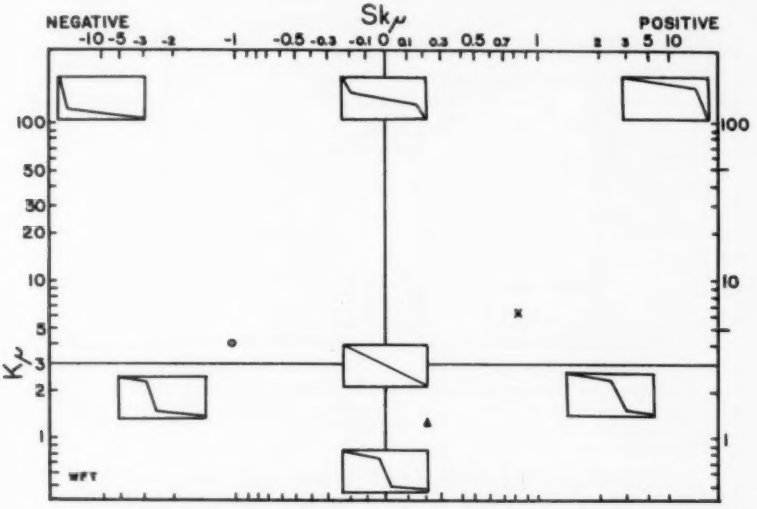



Fig. 2. Representative areas shown on a plot of skewness (on a tangent scale) against kurtosis. Seven idealized curves are shown, including the perfectly symmetrical Gaussian distribution near the center of the figure. The three symbols indicate the plotted positions for the following areas (see fig. 1): circle, Mt. Scott, Okla.; cross, Flagstaff, Ariz.; triangle, Ft. Payne, Ala.

region with elevations ranging from 200 to 600 feet, a point at 400 feet has been noted as 50 percent or 0.50). This has the advantage of facilitating comparison and minimizing the necessity for using the standard deviation; it has the disadvantage of obscuring the absolute depth of dissection. Figure 2 includes the diagrammatic form of typical curves.

An analysis of this kind can be performed on the raw data in much less than an hour for each area to be evaluated. However, it may be advisable, in quick or exploratory work, to use a simpler (but less precise) system. For this purpose, skewness ($Sk_D$ or $Sk_P$) can be plotted against the median

($P_{50} = D_5$), the latter serving as a more-or-less suitable substitute for the mean. The use of a measure of central tendency, in an effort to gain speed, is suggested in view of the fact that there is no suitable alternate measure for kurtosis. The formula for this skewness is quite different from the Sk $\mu$ (based on moments) given above. It can be given in two similar forms:

$$Sk_D = (D_1 + D_9 - 2D_5) / (D_9 - D_1).$$
$$Sk_P = (P_1 + P_{99} - 2P_{50}) / (P_{99} - P_1).$$

These do not yield identical results. The values shown in figure 3 were obtained by using $Sk_D$.

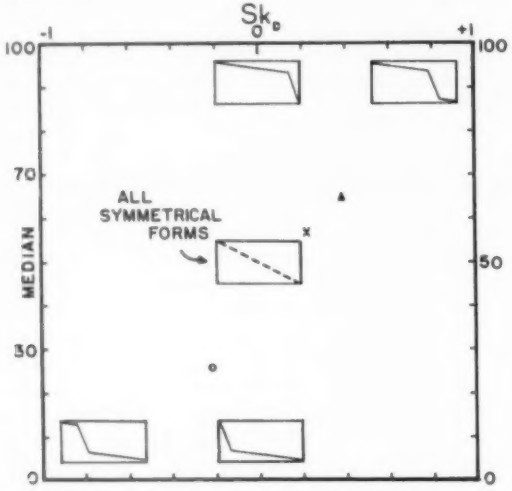



Fig. 3. Representative areas shown on a plot of decile skewness against the median. Five idealized curves are shown, including all symmetrical forms, whether Gaussian or not, near the center of the figure. The three symbols indicate the plotted positions for the following areas (see fig. 1): circle, Mt. Scott, Okla.; cross, Flagstaff, Ariz.; triangle, Ft. Payne, Ala.

Both charts have the disadvantage of indicating featureless early youth, featureless old age, and ideal maturity, in the same way. Recourse may be had, in such circumstances, to the absolute depth of dissection.

The procedures listed here are not difficult to use, and do not require elaborate equipment. Hoel (1947), for example, provides the necessary details in the first 17 pages of his text. However, extra care must be exercised in constructing the array, and calculating the moments, to ensure that the proper sign (eigher positive or negative) is obtained for the skewness.

## References

Hoel, Paul G., 1947, Introduction to mathematical statistics: New York, John Wiley and Sons, Inc., p. 1-17.

Kendall, M. G., 1947, The advanced theory of statistics: London, Charles Griffin and Company, v. 1, p. 137-145.

Tanner, Wiliam F., 1954, Wave-cut erosion surface in the Wichita Mountains, Shale Shaker: Oklahoma City Geol. Soc., v. 5, p. 5, ff.

————, 1958, The zig-zag nature of Type I and Type IV curves: Jour. Sed. Petrology, v. 28, p. 372-375.

## REVIEWS

*Geophysics and the IGY, Proceedings of the Symposium at the opening of the International Geophysical Year*; edited by HUGH ODISHAW and STANLEY RUTTENBERG. P. 210; illus. Washington, D. C., 1958 (American Geophysical Union, Geophysical Monograph No. 2).—Just before the official start of the International Geophysical Year a special symposium was held under the auspices of the United States IGY committee. The purpose was to disseminate information about the activities of American IGY research groups to the scientific community and the public at large. The publication contains thirty papers dealing with all the IGY disciplines, which range from solar and atmospheric studies to oceanography, glaciology and seismology. Many of the papers are by outstanding leaders in research. They are all intended for the non-specialist, but otherwise differ somewhat in emphasis and scope. The subject matter included in various papers overlaps somewhat, but the repetition is not serious.

About a fourth of the volume consists of five papers devoted to rocketry and satellites, the "hardware" that is making possible most of the completely new observations of the IGY. Much of the material on satellites may already be familiar to readers through the many newspaper and magazine accounts that have appeared on the subject. Equally impressive, however, is the role of smaller rockets in probing the upper reaches of the atmosphere. Years of patient work in standardizing and perfecting the controlling and telemetering equipment used with these rockets have produced an extremely valuable and relatively reliable scientific tool.

Another block of papers is devoted to reporting the results of specific investigations completed in advance of the IGY. It is remarkable to see how some fields with an old tradition are suddenly expanding their scope. Geomagnetism is an outstanding example. Exacting observational work during the two forbears of the IGY, the International Polar Expedition of 1888-1889, and the Polar Year of 1933-1934, did a great deal to construct a picture of the magnetic field over the earth and establish the existence of magnetic storms. Suddenly a whole new dimension is being added to the subject through cosmic ray studies and direct observations well out from the Earth's surface. Such programs aimed at determining the complete three-dimensional distribution over the earth of quantities of geophysical interest are a principle goal of the IGY. The large-scale effort required will undoubtedly be rewarded by the discovery of many relationships that tended to be obscured as long as conclusions had to be based on observations relatively isolated with respect to space and time.

Perhaps a third of the papers presented at the symposium contain short reviews of the status of a particular branch of geophysics. Of particular interest are new relationships which cut across traditional distinctions between disciplines. As an example, a paper by Walter Munk of Scripps Institute of Oceanography on one aspect of the IGY program in oceanography suggests that the mean sea level may change considerably with season, something that would be of considerable importance in geodesy. Another paper by Maurice Ewing of Lamont Geophysical Observatory reports that seismologists are finding new information that challenges accepted geological theories of mountain building.

Unfortunately, most of these review papers are accompanied by rather scanty lists of references, which would be inadequate for a reader wishing to learn more about the subject.

For a more comprehensive view of the latest theories and interpretation on a high technical plane, the reader will have to look elsewhere. Nevertheless, these proceedings fulfill an important function by providing a readable account of the actual techniques and problems involved in the American IGY research, with enough background to make its objectives understandable.

KIRK BRYAN, JR.

*Studies on Fossil Vertebrates: Essays presented to D. M. S. Watson;* edited by T. Stanley Westoll. London, 1958 (The Athlone Press, 35*s*; Essential Books, Fairlawn, New Jersey, $5.60).—During the past half century Professor D. M. S. Watson of University College, London, has repeatedly enriched the sciences of evolution and vertebrate phylogeny by his deep insight into anatomical relationships and broad knowledge of geologic history. This fine group of essays, written in his honor by his colleagues and former students, includes penetrating analyses of several knotty problems, useful summaries of various ancient faunas, and philosophical discussions of the evolutionary process.

Three anatomical studies on early fishes by Anatol Heinz, F. R. Parrington, and T. S. Westoll deal largely with the anaspid ostracoderms and present considerable new data on this poorly known order. The perplexing problem of the original environment of vertebrates is analyzed by Errol White who concludes in favor of a marine origin. Other aspects of the paleoecology of fish faunas are discussed by Dorothy Rayner, and the review of Australian fossil vertebrates by E. S. Hills reveals that these consist largely of a fine sequence of fish faunas. Thus half the volume may be said to pertain to paleoichthyology. This is altogether fitting in view of Professor Watson's notable interest in this field. But it is somewhat surprising and disappointing that not a single article on the anatomy of either labyrinthodonts or mammal-like reptiles, two other vertebrate orders to which Watson contributed so much, was included! The two articles on fossil reptiles, by A. S. Romer and E. H. Colbert consider environments and evolution of Permian and Triassic tetrapod assemblages. G. R. De Beer and W. E. Swinton attempt to show by examples from the fossil record that caenogenesis and paedomorphosis rather than recapitulation are characteristic of vertebrate evolution. (The essay by Westoll appears to be based upon the opposite assumption.) James Brough, in a critique of G. G. Simpson's evolutionary studies, challenges the sufficiency of the modern synthetic theory of evolution to account for accelerated evolution at times in the past and to explain the origin of major systematic groups. W. K. Gregory discusses the complex interrelationships of evolutionary factors.

The essays are arranged by the alphabetical order of their author's surnames, which is admirably impartial but also unrelated to their content. Actually, each essay is closely related to certain others, and the group as a whole covers most of Professor Watson's wide field of interest in the anatomy, evolution, function, and environment of ancient vertebrates. They form a most filling tribute to a great scholar.

JOSEPH T. GREGORY

*Sea Shells of Tropical West America*; by A. MYRA KEEN. P. xi, 624, illus. Stanford, 1958 (Stanford University Press, $12.50).—Malacologists all over the country have been waiting impatiently for this work, and now that it is available they will not be disappointed. The book covers shells of all five classes of mollusca found in the Panamic province, which extends from the northern end of the Gulf of California to Colombia. Here, for the first time, the marine shells of this prolific area are brought together in a single volume. More than 1600 species are described, and about 1500 are illustrated. Many photographs are new and many are copies of drawings and photographs from other publications; wherever possible the illustration shows the originally described specimen. In addition, some 70 type specimens are illustrated here for the first time.

The descriptions are brief but adequate. Each species or subspecies is numbered, then comes its name, author, and synonyms, if any; next size, color, distribution, type of shore preferred, and relative abundance are described. There is little or no discussion of shell features that can be clearly observed in the illustrations. An excellent feature is the grouping of the descriptions on the page facing the illustrations, so that the reader always has the text and the picture in front of him. The plates carry no legends, but each shell is numbered to correspond with its description. Ten superb plates in full color add much to the book's attractiveness.

The appendix includes a glossary of malacological terms, keys to the major groups, and a bibliography of 29 pages which lists some 600 titles covering practically everything that has been published about the mollusks of this region. There is an adequate index, and end papers provide maps of the area, with keys to place names and noted collecting localities.

This is a book that will be used by both the specialist and the amateur. With the improvement of roads and travelling accommodations down Mexico way, more and more tourists are vacationing south of the border. Miss Keen's book will be invaluable to those who have a curiosity or interest in the shelled creatures that are to be found along the beaches, and it is an absolute *must* for anyone concerned with the study or classification of the mollusks of North America.

P. A. MORRIS

## ENTOMOLOGICAL ILLUSTRATOR

The Civil Service Commission of Canada invites applications for a position involving the preparation of high quality illustrations of minute insect and insect parts for taxonomic purposes. Candidates must have completed a course of recognized standing in art, have a number of years of related experience, and be capable of training junior staff and developing illustrating techniques.

Salary scale $6210 - $6660 per annum. Excellent provisions for pension, leave and other benefits. For applications write to the Civil Service Commission, Ottawa, Canada, quoting competition number 59-481.

Applications will be received until June 30, 1959.

## ERRATUM

The paper entitled "Lamont Natural Radiocarbon Measurements V", first published in this Journal in January 1959, has been republished in the RADIOCARBON SUPPLEMENT, v. 1, and the authors have requested us to make the following changes as a result of a re-run on sample L-399D (p. 23, both journals). The date of L-399D now reads $9500 \pm 500$; the bone carbonate fraction, $9580 \pm 200$; the untreated fraction, $9150 \pm 230$; the residue fraction, $9500 \pm 500$.

On page 2 of the same paper, end of paragraph 4, "Grotte de la Garenne (14,200)" now reads "Grotte du Renne (15,500)".

VOL. 257 No. 6, 1959

# American Journal of Science

JUNE 1959

## CONTENTS

The Origin of Deformation Lamellae in Quartz
*John M. Christie and C. B. Raleigh* 385

A Metamorphosed Differentiated Sill in Northern Michigan
*Richard W. Bayley* 408

Factors Governing Emplacement of Volcanic Necks ..... *A. R. McBirney* 431

Shell Dunes of the Sonoran Shore ..... *Ronald L. Ives* 449

Examples of Departure from the Gaussian in Geomorphic Analysis
*William F. Tanner* 458

### REVIEWS

Geophysics and the IGY; edited by Hugh Odishaw and Stanley Ruttenberg
*Kirk Bryan, Jr.* 461

Studies on Fossil Vertebrates; edited by T. Stanley Westoll ..... *Joseph T. Gregory* 462

Sea Shells of Tropical West America; by A. Myra Keen ..... *P. A. Morris* 463